The radiative properties of snow at Summit, Greenland

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Under the microscope, I found that snowflakes were miracles of beauty; and it seemed a shame that this beauty should not be seen and appreciated by others. Every crystal was a masterpiece of design and no one design was ever repeated. When a snowflake melted, that design was forever lost. Just that much beauty was gone, without leaving any record behind. 

Wilson A. Bentley, *The Snowflake Man* (1865 – 1931)
Abstract

The amount of solar radiation reflected at the earth surface is one of the primary parameters controlling the surface radiation balance which determines the heat and water exchange processes. Snow and ice cover a seasonally varying area of up to 15% on this planet. Therefore a profound knowledge of the snow radiative processes is needed to calculate the planetary radiation balance. Among all surfaces snow in particular shows high reflectance and complex absorption features. However, snow properties influencing the magnitude of reflectance and absorption are not constant over time. For example the anisotropic reflectance of a snow cover over the hemisphere varies with the solar incident angle. Presently, most general circulation models use a fixed snow albedo value over the globe or they apply a simple parametrization to calculate the surface reflectance based on two broadband radiation ranges (e.g. visible and near infrared). In remote sensing on the other hand, the broadband hemispherical albedo has to be extrapolated from a discrete and small number of confined spectral bands, and usually measured from a single viewing direction. Both applications generally fail to resolve the fine spectral and directional reflectance properties of snow, which in fact determine the broadband albedo.

As a first step in mitigating this unsatisfying situation and improving the understanding of the snow radiative properties, the Summit Environmental Observatory on the ice sheet in the dry snow zone of Greenland has been chosen to carry out detailed studies on absorption, transmittance and reflectance properties of snow. The research focuses on the penetration depth of solar radiation into the snow and on the directional reflectance of solar radiation back to the hemisphere. The specially developed measuring instrumentation is presented which was used during three field campaigns, in 2001, 2004 and 2005.

In the first part of this work the climatological conditions at the measuring site are shown. The relevant snow radiative processes are then summarized. A detailed literature review demonstrates the complexity of the involved problem and shows the present state of research. The experiments of 2001, investigating on the penetration and extinction of solar radiation within the snow cover, confirm the selective spectral absorption of snow and the exponential pattern of the decreasing intensity with depth. Results of this study show that a large portion of the absorbed radiation does not penetrate deep into the snow and it is assumed to be removed in the form of latent heat flux. However, processes at the snow-atmosphere interface are still not well understood.

The second part of this work presents the new IACETH field Gonio-Spectrometer developed at the Institute for Atmospheric and Climate Science. It measures directional reflected radiation with high spectral and spatial resolution and a hemispherical directional reflectance factor can be calculated. The goniometer is fully automatic and the two robotic arms are controlled with step motors. A spectrometer equipped with a 3° foreoptic is used for taking spectrums of the reflected radiation in the range of 350 to
1050 nm. The time needed to collect one complete hemispherical dataset (15° resolution in both view zenith and view azimuth angle) is 11 minutes, corresponding to a change of less than 4° in solar zenith and azimuth angle.

The datasets collected with the IACETH Gonio-Spectrometer are presented and used in the third part of this work. The Hemispherical Directional Reflectance Factor (HDRF) is used as a means to exploit the directional reflectance properties of snow. The observations were carried out for two prevailing snow surface types: a smooth surface with windbroken small snow grains and a surface covered with rime causing a higher surface roughness. The HDRF distribution was nearly isotropic at noon. It varied with increasing solar zenith angle, resulting in a strong forward scattering peak. Smooth surfaces exhibited stronger forward scattering than surfaces covered with rime. At a solar zenith of 85°, an HDRF of ~13 was observed in the forward scattering direction for λ=900 nm. Spectral albedos were then calculated by interpolating the HDRF datasets on a 2° grid and integrating individual wavelengths. Spectral albedos showed variations depending on the solar illumination geometry and were sensitive to snow surface properties. Broadband albedos were calculated by integrating the spectral albedos over all wavelengths. The broadband albedos derived from directional measurements reproduced the diurnal pattern measured with two back-to-back broadband pyranometers.

Finally, a laboratory experiment allowed to carry out investigations on the spectral snow reflectance. Using tomography, snow samples taken in the area of Davos were analyzed in detail with respect to their small scale structure and the specific surface area (SSA). The SSA correlates well with the measured reflectance. However, low values in the visible range which might be caused by impurities, could not be explained since in this study impurities have not been measured.

In summary, this study shows that the snow radiative processes are highly dependent on small and medium scale snow physical characteristics. It exemplifies that snow radiative properties can be determined by use of a consistent measurement framework which integrates surface radiative processes, snow physics and snow morphology. Ultimately snow chemistry, surface roughness and further snow physical parameters should be concurrently measured as part of a future field experiment. This study presents a foundation for further investigations of this kind with the ultimate aim of incorporating the radiative characteristics of snow into energy balance studies and climate models with complex topography.
Zusammenfassung


Sonnenzenit und -azimutwinkels von weniger als 4°.


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List of Acronyms and Abbreviations

"Aput": A Tribute to the Inuit Language

Acknowledgments

Curriculum Vitae
Chapter 1

Introduction

"Where does the white go when snow melts?"
It’s a decent question to start with.
And the more you think about it,
the more questions it spins off.
Like, why is snow white to start with?

1.1 Solar radiation, snow and climate

The surface reflective properties determine how much of the incoming radiation is reflected back to space. This ratio is either called reflectance or albedo. Albedo is a Latin word and means whiteness. It is used to describe how bright or white something is. In general darker surfaces have a lower albedo and absorb more solar energy than lighter surfaces. Among all surface cover types of the Earth, snow is the brightest one and clean snow shows albedo values close to one. In a more global view polar regions with their extended snow cover are cold since they receive less solar radiation and also reflect a great portion of the incoming radiation back to the atmosphere and thus have a lower amount of net energy available to heat the surface.

It has been known for a long time that snow exerts a strong influence on small scale (Miller, 1956; Schlatter, 1972) but also large scale climate patterns (e.g. Budyko, 1969; Konzelmann and Ohmura, 1995; Hall, 2004). In the context of global climate change and sea level rise the important role of the polar regions is widely acknowledged. Polar regions show a high sensitivity to small changes in radiative forcing partly as a consequence of the ice and snow surface albedo feedback mechanism (Budyko, 1969; Hall, 2004; Zwally et al., 2002; Winton, 2006). The classical example of this albedo effect is the snow-temperature feedback: if a snow covered area warms (due to e.g. a positive radiative forcing) the snow melts and the albedo decreases; more sunlight is absorbed by the surface which results in a even stronger surface warming than it would be observed without the involved snow-albedo feedback. The reverse process can also occur: if snow forms, a cooling cycle happens. Hansen and Nazarenko (2004) simulated the effect of soot on the snow and ice albedo feedback. They found plausible estimate of 0.3 W/m² for the climate forcing in the northern hemisphere.

The energy and mass balance is an important starting point in understanding the response of the ice sheet and the resulting sea-level change to climatic variations (Alley et al., 2005; Zwally et al., 2002; Konzelmann, 1994; Ohmura et al., 1996).
Seasonal snow cover, the largest component of the cryosphere in terms of the area, varies seasonally and covers up to 15% (Ohmura, 2006) of the Earth’s total land surface. Detailed studies on the energy balance of snow covers have been carried out by e.g. Liljequist (1956); Colbeck et al. (1975); Ohmura (1980); Kane et al. (1991); Brandt and Warren (1993). The energy balance of a snow cover is a variable dynamic surface boundary condition in climate models. Thus, understanding of global and regional climate variability and trends requires that we monitor the temporal and spatial variability of the snow covered area.

Snow cover extent and snow albedo are interactive Earth system components in climate models so the accuracy of their parametrization can have a large effect onto the simulated climate (Sellers et al., 1995; Zhou et al., 2003; Hanna et al., 2005). Climate models are currently being used as tools to understand past climatic events on Earth and they’re extensively being used to prognose the expected future climate changes (IPCC, 2001). It is therefore a requirement for climate model simulations that snow radiative processes are well understood to a degree where they can be implemented as mechanistic and physically-based formulations in these models. Due to a lack of global measurements of snow radiative properties satellite-derived snow datasets are replacing simplified look-up table snow albedo values in climate models. An extensive set of Earth observing platforms (e.g. NASA EOS or ESA Envisat) also serves the international climate research community with high quality and spatially continuous snow data. It has recently become a policy requirement to monitor the Earth’s physical, chemical and biological processes on a operational schedule (GEOSS - Global Earth Observation System of System). The development of snow radiative parameters for both remote sensing and climate modeling however extensively depends on ground measurements. They are a requirement for parameter estimation and validation (Justice et al., 2000).

1.2 Reflectance characteristics of snow

As already mentioned snow owns a special role among all land cover types. It reflects a great part of the incoming shortwave radiation and shows a characteristic spectral reflectance signature compared to other surfaces such as sand, deciduous woodland or black loam (Fig. 1.1). The reflectance of clean snow is higher than 0.8 in the visible and near infrared region and decreases strongly at larger wavelengths. However, due to changing solar illumination geometry, snow metamorphism and the deposition of impurities snow albedo exhibits temporal and spatial variations (Carroll and Fitch, 1981; Ohmura, 1980; Yamanouchi, 1983; McGuffie and Henderson-Sellers, 1985). Satellite remote sensing of snow offers a powerful tool to quantitatively examine the physical properties of snow (snow area extent, reflectance, grain size, snow melt) in cold regions, where traditionally such measurements are difficult and time-consuming (Nolin and Dozier, 2000). With the knowledge of characteristic spectral signatures the physical properties of different types of snow surface can be distinguished (Grenfell and Maykut, 1977; Painter et al., 1998; Li et al., 2001; Green et al., 2002; Aoki et al., 2003).

In recent years satellites have been extensively used to monitor the snow area extent on the Earth (Robinson et al., 1993). As an example Figure 1.2 illustrates the snow area extent for July and January 2004, respectively. The images are based on TERRA MODIS land surface reflectance data.
1.3 State of the research related to snow reflectance

Optical properties of snow are well explored. But the variety of physical snow characteristics (grain size, grain shape, snow cover density, snow cover stratigraphy, water content, contaminations) and the temporal changes of these characteristics are complex to model.

Many studies have investigated snow albedo by conducting measurements at one site for a short time period. Only few networks exist that are measuring snow albedo operationally. On the ice sheet of Greenland the Institute for Atmospheric and Climate Science measures all radiation components according to the high standards of the Baseline Surface Radiation Network (BSRN) (Ohmura et al., 1998). The BSRN network is a globally distributed radiation measuring network which provides a continuous and high quality data record. The Greenland Climate Network (Steffen and Box, 2001) consisting of twenty automatic weather stations distributed over the ice sheet of Greenland also operationally collects meteorological datasets.

A large number of studies (e.g. Hall et al., 1992; Dozier, 1989; Knap and Oerlemans, 1996; Painter et al., 1998; Han et al., 1999; Jin and Simpson, 1999, 2001; Tanikawa et al., 2002; Stroeve and Nolin, 2002; Stroeve et al., 2005) derived snow area coverage, physical snow properties and radiation conditions by using satellite datasets. With the advent of satellite sensors that offer high spectral, spatial and temporal resolution as well as good in-flight calibration (such as MODIS on board NASA’s TERRA and AQUA, Hyperion on board NASA’s EO-1 or ESA’s MERIS on board ENVISAT) the remotely sensed data became interesting for the snow community to investigate snow cover extent and snow cover properties. The National Snow and Ice Data Center in Boulder, CO (USA) hosts a large number of surface-based and remotely-sensed snow datasets (http://nsidc.org).

![Modeled spectral signatures of fine grained snow, coarse grained snow, white dune sand, deciduous woodland and black loam (source: SBDart, Ricchiazi et al. (1998)).]
Fig. 1.2: TERRA MODIS land surface reflectance data for July (top) and January (bottom) illustrating the seasonal variation of the snow covered area on the northern hemisphere (Stöckli et al., 2006).
1.4 Aims and outline of this study

The global climatology group of the Institute for Atmospheric and Climate Science at the Swiss Federal Institute of Technology (IAC ETH) is carrying out a research program which investigates climatic interactions between the boundary layer and the ice sheet. Since the year of 2000 all the components of the radiation balance are operationally measured at the Greenland Summit Environmental Observatory (72° 35' N, 34° 30' W, 3203 m a.s.l).

The present study is an integrative part of the above mentioned measuring campaign. The overall goal is to provide a better understanding of the reflectance characteristics of a dry snow cover with a special focus on its directional reflectance properties.

Chapter 2 presents a summary of the scientific IAC ETH measuring program and provides an overview on the climatological conditions at Summit during the field seasons of 2004 and 2005. The theory on snow radiative processes is given in Chapter 3. An important step in this work was the development of a new Gonio-Spectrometer. The description of the instrumentation is given in Bourgeois et al. (2006b). The paper is included in this thesis as Chapter 4. With the Gonio-Spectrometer, data was collected during two field seasons in 2004 and 2005. The data analysis focused on the Hemispherical Directional Reflectance Factor (HDRF), and the results are summarized in Bourgeois et al. (2006a), Chapter 5 of this thesis.

During the winter of 2005/2006, laboratory investigations of the spectral albedo for various snow types were carried out at the Swiss Federal Institute for Snow and Avalanche Research Davos (SLF). Preliminary results of these investigations are included in Appendix A.
Chapter 2

The Greenland Summit Environmental Observatory

Grönland gehört nicht nur für seine eigenen Bewohner, sondern auch für uns europäische Besucher zu jenen Ländern, die Heimweh machen. Wir können beim Abschied nicht denken, dass wir das alles nie mehr sehen sollen.

Alfred de Quervain (1879 – 1927)

2.1 Measuring site

The study site, Summit (72° 35’ N, 34° 30’ W, 3203 m a.s.l.) is located in the dry snow zone on the top of the ice sheet. The surface area of the dry snow zone on the Greenland ice sheet was estimated to be $683 \times 10^3$ km$^2$ by Benson (1962). This corresponds to 40% of the total ice sheet surface area. Newer studies show, the area with dry snow has decreased in recent years. However, the climate at Summit can still be regarded as representative for a major part of the dry snow zone of the Greenland ice sheet. The significance of the Greenland ice sheet for the climate in the Northern Hemisphere and the global sea-level, has given rise to extensive scientific investigations on the ice sheet (Putnins, 1970). Past climate is studied with the interpretation of ice cores. Observations of the structure of the atmosphere above the ice sheet and the physical processes at the surface and in the snow cover are necessary to understand the energy and mass balance of the ice sheet.

Summit was established in 1988 by the United States National Foundation for the Greenland Ice Sheet Project II (GISP II). In 1993 the GISP II drilling project completed the retrieval of a 3054 meter ice core. In order to enable a more comprehensive analysis of the ice core record, intensive field campaigns were conducted at the site to characterize the atmosphere-to-firn dynamics. In Fig. 2.1 a sketch of the Summit station is drawn. The locations of the IACETH instruments are marked in green (general measurements) and red (important for this work).

2.2 IACETH measuring program

The IACETH (former Dept. of Geogr.) has a long tradition of research in the Arctic and on the Greenland ice sheet. In 1990 and 1991 scientific expeditions at the mean equilibrium line were carried. At the site of today’s Swiss Camp energy and mass balance studies of
the ice sheet were performed (Ohmura et al., 1991, 1992). At the Summit Environmental Observatory (Ohmura, 2001), year-round measurements of all the components of the energy balance are carried out since 2001 by the IAC ETH. The activities focus on the energy and mass balance of the ice sheet and the investigation of the planetary boundary layer in the dry snow zone. These include measurements of the radiative fluxes such as direct solar radiation, diffuse sky radiation, global radiation, shortwave reflected radiation, longwave incoming radiation, longwave outgoing radiation and net radiation as well as micrometeorological measurements needed to compute the turbulent exchange of sensible and latent heat. The albedo can be determined from the measurements of outgoing and incoming shortwave radiation fluxes. The conditions of the planetary boundary layer are investigated with an instrumented 50 m tower, including profiles of temperature, wind speed and relative humidity. Furthermore, snow temperature is recorded and irregular snowpit work is performed to determine snow density and stratigraphy.

The radiation conditions and energy balances at Summit have been studied extensively and presented by Hoch (2005). The turbulence characteristics of the atmospheric boundary layer are the subject of the thesis of Schelander (2006). The extinction processes and the radiation fluxes of solar radiation within the snow cover have been investigated and are presented in Bourgeois (2002) and Meirold-Mautner (2004). The snow stratigraphy and energy balance has been investigated by Dadic et al. (2004). To complement previous studies the present work focuses on the spectral and directional surface reflectance at Summit.

2.3 Climatological conditions at Summit during the 2004 and 2005 campaigns

Figures 2.2 and 2.3 show the time development of the following climatological parameters during the field seasons 2004 and 2005: the global radiation, albedo, direct radiation, 2 m temperature and 2 m wind speed. In 2004 the measuring period lasted 40 days starting on 27 May, in 2005 it lasted 35 day starting on 18 June. The mean calculated albedo in 2004 was 0.828 (standard deviation: ±0.021) and 0.837 (±0.029) in 2005. Only values where the solar zenith angle was smaller 70° have been considered. The mean 2 m temperature in 2004 was -15.76°C (±4.90), the mean temperature in 2005 was -12.64°C (±4.30). The mean 2 m wind speed in 2004 was 2.45 m/s (±1.69), the mean wind speed in 2005 was 3.50 m/s (±2.23). Clear sky conditions were required for the directional reflectance measurements. In 2004 the clear sky periods were distributed equally over the whole season while in 2005 favorable periods lasted several days in the beginning and at the end of the season.
2.3. Climatological conditions at Summit during the 2004 and 2005 campaigns

Fig. 2.1: Sketch of the Summit Environmental Observatory.
Fig. 2.2: Micrometeorological measurement time-series of the 2004 campaign: Global radiation (sum of the cosine corrected direct radiation and the diffuse radiation), albedo, 2m temperature and 2 m windspeed.
2.3. Climatological conditions at Summit during the 2004 and 2005 campaigns

**Fig. 2.3:** Micrometeorological measurement time-series of the 2005 campaign: Global radiation (sum of the cosine corrected direct radiation and the diffuse radiation), albedo, 2 m temperature and 2 m windspeed.
Chapter 3

Basic Concepts of Snow Radiative Processes

There is no material that under normal conditions displays the bewildering complexities found in snow.
M. Mellor

3.1 Snow optical properties

From the physical point snow is a composite of ice crystals, water and air. The physical properties of both materials are well explored. However, snow exhibits a completely different behavior compared to ice crystals and air. While air and pure ice are translucent for the human eye, pure snow is white. It reflects most of the incoming radiation in the visible range of the spectrum. In the near- and far-infrared wavelength region snow becomes grey and even black and absorbs most of the incoming radiation. Since the visible part of the solar radiation contains the largest portion of the emitted energy, a beam of sunlight entering a snow bank is quickly scattered by ice crystals and air pockets and is scattered back into the atmosphere. As a result snow appears to be white for the human eye.

Snow optical properties have been studied in theory and/or with experiments: some of the early works include Sauberer (1938), Thams (1938), Dunkle and Bevans (1956), Liljequist (1956), Giddings and Lachapelle (1961), Mellor (1966), Bohren and Barkstrom (1974), Grenfell and Maykut (1977). All the studies agree that the significant parameters influencing the radiative processes of snow are the snow grain size, snow grain shape, snow depth, snow density and the content of impurities in the snow.

As radiation enters a snow pack it is transmitted, absorbed or scattered. The dominant optical property that causes the spectral variation of the snow reflectance is the spectral variation of the imaginary part of the complex refractive index of ice. It increases five orders of magnitude \((10^{-9} - 10^{-4})\) between 350 and 1000 nm (Warren, 1984). The real part stays nearly constant \((1.32 - 1.30)\) for the same wavelength range. The complex part is responsible for the absorption. This explains the transparency of ice in the visible range and the increasing absorption in the near infrared. However, already few absorbing impurities in the snow / ice can significantly decrease the visible reflectance (Warren and Wiscombe, 1980).

Warren (1982) presents a comprehensive overview on the snow optical properties: an
3. Basic Concepts of Snow Radiative Processes

Important parameter is the single scattering albedo. It depends on the extinction efficiency and the scattering phase function. The single scattering properties are often described with the Mie-theory. However, a snow cover consisting of ice crystals and air allows for multiple scattering. Two approaches have been used to model the multiple scattering: the delta Eddington approximation, essentially a two-stream approximation, and the discrete ordinates method (Stamnes et al., 1988) solving the radiative transfer equations.

With improving measuring technology, spectral and directional measurements can be carried out and the above snow radiation models can be tested and improved. Furthermore, increasing computing resources now allow to simulate complex processes in radiative transfer models.

3.2 Broadband reflectance of snow

Energy balance studies require knowledge of the hemispherically integrated broadband albedo. This corresponds to the total reflectance of the shortwave solar radiation. The equation of the broadband reflectance or albedo, $\alpha$, reads:

$$\alpha = \frac{E_{out}}{E_{in}},$$

where $E_{out}$ denotes the total upward directed radiative flux and $E_{in}$ the downward radiative flux of shortwave radiation. Usually the wavelength range is defined from 280 nm to 4 $\mu$m. First studies of the solar reflectance and transmittance of snow were conducted by Thams (1938) and Sauberer (1938). A number of studies followed, taking advantage of improving measuring technology. Ohmura (1980) reports on the daily cycle of the broadband albedo for dry and for wet snow, on clear and on cloudy sky conditions. McGuffie and Henderson-Sellers (1985) discuss the diurnal hysteresis of snow albedo and explain it with either snow metamorphosis or surface irregularities. Stroeve et al. (1997) compared AVHRR derived albedo data with in situ measurements on the Greenland ice sheet. Wendler and Kelley (1988) carried out detailed albedo measurements in the dry snow zone of Antarctica and demonstrate the influence of cloud cover and snow age on the reflectance. Sicart et al. (2001) conducted clear-sky albedo measurements on a sloping glacier and present a correction for the reflectance measurements on tilted surfaces. Aoki et al. (2003) report on the influence of snow physical parameters on the shortwave broadband albedo and use a radiative transfer model to simulate their measurements. Pirazzini (2004) compares surface albedo data from several Antarctic sites to determine its spatial and temporal variability.

In general the broadband albedo is high for fresh snow and decreases with snow age. The main reason for this behavior is the growth of the snow grains due to snow metamorphosis. The daily variation of the broadband albedo is a result of the spectral variation of the incoming irradiance in the course of the day. The albedo shows the highest values when the sun is low or when the sky is overcast. This is attributable to the predominantly blue to green spectral composition of sky-diffuse radiation for which the snow cover possesses the largest reflectivity (see next section) and the mirror-like reflection. Figure 3.1 shows the albedo at Summit for four overcast days and four clear sky days together with the direct component of the solar radiation. The values of the albedos range from 0.79 to 0.91. The values are up to 10% higher when the sky is overcast. Furthermore on clear sky days a distinct daily cycle is visible.
3.3 Spectral reflectance of snow

Knowledge of the spectral signature of snow is important for at least three reasons:

- First, to determine an accurate heat balance of a snow cover the spectral reflectance is necessary.

- Second, satellite remote sensing of snow properties will become more important with the increasing number of space-based observation platforms in the future. However, most instruments on satellites measure a limited number of spectral bands. To determine a broadband surface albedo a narrow to broadband conversion has to be applied to the remotely sensed data. Knowledge of the spectral reflectance pattern is required for such a conversion scheme. Narrow to broadband conversion schemes for snow have been put forward by Liang et al. (2003) and Greuell and Oerlemans (2004).

- Third, the growth of the vegetation underneath a snow cover depends on the radiation penetrating through the snow. The knowledge of the penetrating PAR is therefore essential for modeling the vegetation processes under a snow cover.
The equation for the spectral radiation reads:

\[ \alpha_{\lambda} = \frac{E_{\lambda,\text{out}}}{E_{\lambda,\text{in}}} \tag{3.2} \]

where \( \lambda \) signifies the spectral dependency. Spectroscopic reflectance studies on Antarctica snow are reported in Kuhn and Siogas (1978). Choudhury and Chang (1979) present a parametrization of the spectral albedo using the single scattering albedo and the fraction of energy scattered in the backward direction. In their study calculated reflectances for different snow types were in good agreement with the measurements. Wiscombe and Warren (1980) presenting *A Model for the Spectral Albedo of Snow I: Pure snow* and Warren and Wiscombe (1980) and *A Model for the Spectral Albedo of Snow II: Snow Containing Atmospheric Aerosols* belong to the most often referenced publications in the field of snow and spectral reflectance. The model of Wiscombe and Warren uses the 'delta Eddington' approximation for multiple scattering and the Mie theory for single scattering. They find that the spectral albedo is dependent on the effective grain size, the solar zenith angle, the snow pack thickness, and the ratio of diffuse to direct radiation. Figure 3.2 is taken from the study Wiscombe and Warren (1980) and illustrates the variation of the spectral albedo with grain size (3.2a) and solar zenith angle (3.2b). Choudhury et al. (1981) present a parameterizations for the reflectance of snow containing impurities. Grenfell et al. (1981) measure the spectral reflectance of an alpine snowpack for fresh snow and for metamorphosed and melted snow. Davis et al. (1993) model the temporal changes of the spectral signature of snow and validate the results with laboratory measurements. Green et al. (2002) present two spectral snow reflectance models which account for the grain size and the liquid water fraction with the final goal to derive grain sizes from spectral signatures. Zhou et al. (2003) model the effects of vertical inhomogeneity on snow spectral albedo.

In this study we made laboratory measurements of the spectral snow albedo for the wavelength range 350 – 1050 nm. Five different snow types were analyzed. The data is
analyzed with respect to its specific surface area (SSA) and it will be compared to a beam tracing model. Preliminary results are summarized in Appendix A.

3.4 Directional reflectance of snow, BRDF, HDRF and HCRF

One of the main characteristics of snow surfaces is its highly anisotropic reflectance back to the hemisphere depending on the illumination and reflectance geometry. Other factors influencing the anisotropy are the snow physical properties like grain size, grain shape, density, stratigraphy, snow depth, impurities in the snow and the wavelength. The directional reflectance can therefore not be expressed as a constant parameter; it is rather a function over the hemisphere. The anisotropic reflectance is of prime interest for satellite measurements. Satellite sensors with a narrow field-of-view generally see the surface from a single view angle (special sensors like MISR have a few discrete angles). In the literature the nomenclature to describe directional reflectance properties is not consistent and has often been misapplied. Some studies also use self-created factors and indices. The first nomenclature for directional reflectance has been published by Nicodemus et al. (1977). However, only the bidirectional reflectance is mentioned there which – strictly spoken – only considers directional incoming and outgoing radiation. In a later study Martonchik et al. (2000) reviewed the whole terminology for use in remote sensing. The most recent work from Schaepman et al. (2006) elaborates all details and naming conventions in the field of directional radiation with case studies. The definitions for the bidirectional reflectance distribution function (BRDF), the hemispherical directional reflectance factor (HDRF) and the hemispherical conical reflectance factor (HCRF) are given here according to Schaepman et al. (2006).

The BRDF describes the scattering of a parallel beam of incident light from one direction in the hemisphere into another direction in the hemisphere:

\[
\text{BRDF}(\theta_1, \phi_1; \theta_r, \phi_r; \lambda) = \frac{dL_r(\theta_1, \phi_1; \theta_r, \phi_r; \lambda)}{dE_i(\theta_1, \phi_1; \theta_r, \phi_r; \lambda)} \text{ [sr}^{-1}] , \tag{3.3}
\]

where L stands for radiance, E for irradiance, the subscript i for the incoming direction, the subscript r for the reflected direction, \(\theta\) for the zenith angle and \(\phi\) for the azimuth angle.

The HDRF is given by the ratio of the reflected radiant flux \(\Phi\) from the surface area \(dA\) to the reflected radiant flux from an ideal (id) and diffuse surface of the same area \(dA\) under identical view geometry. The irradiance is assumed from the entire hemisphere:

\[
\text{HDRF}(\theta_1, \phi_1, 2\pi; \theta_r, \phi_r; \lambda) = \frac{d\Phi_r(\theta_1, \phi_1, 2\pi; \theta_r, \phi_r; \lambda)}{d\Phi_{id}^r(\theta_1, \phi_1; 2\pi, \lambda)} \text{ [-]} . \tag{3.4}
\]

Finally the equation of the HCRF, the most often measured quantity in the field reads:

\[
\text{HCRF}(\theta_1, \phi_1, 2\pi; \theta_r, \phi_r, \omega_r; \lambda) = \frac{\int_\omega \int_{2\pi} \text{BRDF} \cdot L_i(\theta_1, \phi_1) d\Omega_i d\Omega_r}{(\Omega_r/\pi) \int_{2\pi} \int_{2\pi} L_i(\theta_1, \phi_1) d\Omega_i} \text{ [-]} , \tag{3.5}
\]

where \(\omega\) is the solid angle subtended by the sensor foreoptic and

\[
\Omega = \int d\Omega = \int \cos \theta d\omega = \int \int \cos \theta \sin \theta d\theta d\phi
\]
3. Basic Concepts of Snow Radiative Processes

is the projected solid angle of the cone.

Dirmhirn and Eaton (1975) were among the first measuring the angular distribution of reflected shortwave radiation using a radiometer with a small aperture. A number of theoretical and experimental works followed measuring various snow surfaces and various spectral ranges at various illumination angles. Choudhury and Chang (1981a) and Choudhury and Chang (1981b) discuss the theory of the angular and spectral variation of snow reflectance and compare the theory with a few observations. Kuhn (1985) presents first detailed measurements of the HDRF for a number of solar zenith angles, for different wavelengths and for different snow types. They observe a peak in the azimuth direction up to 60° for the forward scattering direction. Two years later, Steffen (1987) reports on directional measurements for 500 – 600 nm, at solar zenith angles ranging from 28° to 85° and for three different snow types. He uses the expression anisotropic reflectance factor for what is now commonly called HDRF. In this study a first attempt was made to compare the measured field data to satellite data. Hall et al. (1992) measures spectral albedo and also few off-nadir reflectances. The difficulties of comparing surface based data to satellite data is subsequently discussed in that study. Receiving remotely sensed data from homogeneous pixels was then not possible due to the small spatial resolution of past satellite sensors. Grenfell et al. (1994) measure BRDF from a 23 meter tower and discuss possible influences of sastrugis on the directional reflectance. Steffen (1997) reports on experiments conducted on the Greenland ice sheet measuring the anisotropic reflectance with high spectral resolution for different snow surfaces and at different solar zenith angles. Besides of the well known forward scattering peak he observes a backward scattering peak. Sergent et al. (1998) presents experimental results from Hemispherical-directional reflectance measurements and makes comparisons with the adding-doubling method. In a number of publications Leroux and Fily (1998); Leroux et al. (1998, 1999) discuss on polarized directional reflectance and measurements are compared with model results based on the radiative transfer theory also using the adding-doubling method. Warren et al. (1998) investigates the effect of surface roughness (mainly sastrugi) on the directional reflectance at South Pole. The experiments are again conducted from a 22 m tower and for the wavelength 600 nm, 660 nm and 900 nm. Aoki et al. (2000) study the effect of snow physical parameters on the spectral albedo and on the directional reflectance and compare the results with two kinds of theoretical models: one uses the Mie theory and the other the Heney-Greenstein phase function to calculate the single-scattering parameters. Li and Zhou (2004) presents field measurements of directional reflectance measurements of snow-covered sea ice and simulate the pattern with the aid of snow physical data and radiative transfer models. Painter and Dozier (2004a) reports on HDRF measurements collected with the Automated Spectro-Goniometer and compares the data with modeling results from DISORT. Kokhanovsky et al. (2005) compares measurements to a snow optical model where the snow grains are represented as fractal particles. The experiments of Peltoniemi et al. (2005) show the effect of snow grain size and snow grain shape on the snow directional reflectance.

As elaborated in the last paragraph, a number of studies have investigated the directional reflectance properties using different measuring techniques and modeling approaches. Comparing the results of different studies is difficult since the terminology used is often not the same and also the description of the snow conditions varies from one study to the next. One result is common to all studies: the increasing scattering in the forward direction with increasing solar zenith angle. As an example Fig. 3.3 illustrates the HDRF
measured at Summit on two different solar zenith angles. More on this in Chapter 5.

19. July 2005 \( \lambda = 700 \text{ nm} \)

\[ \theta = 51^\circ \]

\[ \theta = 74^\circ \]

backward scattering      forward scattering
1.0
1.0
1.5
2.0
2.0
0.9
1.1
0.8

Fig. 3.3: Examples of typical patterns of the HDRF for a moderate solar zenith angle of 51° and a large solar zenith angle of 74°.

3.5 Solar radiation penetrating in the snow cover

Specific investigations of the extinction of solar radiation in the dry snow cover of Summit were carried out in 2001 and 2004. Results are reported in Bourgeois (2002) and Meirold-Mautner (2004), respectively. The overall picture that emerges from the 2001 experiments is summarized here.

Two sets of instruments were utilized:

a) Sensors composed of silicon photovoltaic detectors with two interference filters, and diffusors. Down-welling and up-welling radiation fluxes in the range of 400 to 700 nm were measured at three levels: above the surface, at -6 cm and at -12 cm. A timeseries for the period of 10 days was recorded.

b) Incoming radiation flux was measured with a spectroradiometer \([350–2500 \text{ nm}]\). The sensor of the spectroradiometer was placed consecutively at different heights (-0.8 cm to -62 cm) below the snow surface. Measurement were carried out on selected days.

A two-stream model was applied on the datasets based on Schuster (1905) and exemplified in detail for snow by Ohmura (1980). The results were analyzed towards the observed snow conditions. Absorption and scattering coefficients were calculated from observations of extinction coefficients.
3.5.1 Broadband radiation and extinction in the snow in the range of 400 to 700 nm

In Figure 3.5 the calculated extinction coefficients, $\kappa$, are shown together with the global radiation. The evaluation of the data was only made for solar zenith angles smaller than 70°.

The mean value for $\kappa$ was 0.071(±0.003) for down-welling radiation and 0.065(±0.002) for up-welling radiation. According to the theory both values should be the same if radiation is measured over all wavelengths in a homogeneous snow layer. However, the extinction coefficient of the downward radiation was higher. There are two possible explanations. First, the measuring range is restricted to 400 - 700 nm. The percentage of radiation above 700 nm is smaller for downward radiation than for upward radiation. The spectral
composition of the solar radiance changes when penetrating the snow, with a stronger depletion for the wavelengths longer than 1000 nm. The percentage of upward radiation between 400 and 700 nm is therefore higher than for downward radiation in the same range. Second possible explanation is a self-shading effect. A sensor facing downward blocks the vertically incident beams. Radiation reaching the top of the sensor is either reflected or absorbed by the sensor. This leads to a change in the intensity, direction and spectral composition of the subsurface radiation. Another remarkable characteristic in the evaluated extinction coefficients was the opposed trend with time for the up- and downward series. The extinction of the downward flux was maximal at noon, whereas, at the same time, the extinction of the upward flux was minimal.

3.5.1. a Absorption- and scattering coefficients

Absorption (k) and scattering (r) coefficients for the range of 400 to 700 nm were calculated and are displayed in Figure 3.6.

![Figure 3.6: Time series of absorption (k) and scattering (r) coefficients. The spectral range is 400 to 700 nm.](image)

According to the theory (Ohmura, 1980) both coefficients depend strongly on the reflectivity, \( r_s \), in the snow. For a homogeneous snow cover \( r_s \) (reflectivity within the snow) should be equal to \( r_{s0} \) (reflectivity at the lower side of the snow surface). Values of the total reflectivity, \( \alpha \), and \( r_s \) for the two levels are presented in Figure 3.7 for the range of 400 to 700 nm.

Due to the selective extinction properties of the snow, the spectral composition of downward and upward radiation differs, and \( r_s \) varies if integrated measurements are considered. A clear temporal variation can be seen in the data. Furthermore, the instruments disturb the radiation field by screening radiation from above and below. Liljequist (1956) made experiments and increased the size of his instruments to determine the screening effect. He proposed a reflectivity in the snow, \( r_s \), of 0.40. Konzelmann (1994) measured values between 0.4 and 0.9. He assumed an underestimation in the measurements of the upward directed flux of 15%. In this study \( r_s \) is found to be between 0.5 and 0.6 for the range of 400 to 700 nm. An average value of 0.54 was assumed for the calculations of \( k \).
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Fig. 3.7: Total reflectivity, $\alpha$, of the snow cover and reflectivity, $r_s$, in the snow for two levels (-6 cm and -12 cm). The spectral range is 400 to 700 nm.

and $r$. Table 3.1 shows the mean values for extinction, $\kappa$, absorption, $k$, and scattering, $r$, coefficients.

<table>
<thead>
<tr>
<th></th>
<th>downward</th>
<th>upward</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\kappa$ [cm$^{-1}$]</td>
<td>0.071(±0.003)</td>
<td>0.065(±0.003)</td>
</tr>
<tr>
<td>$k$ [cm$^{-1}$]</td>
<td>0.020(±0.001)</td>
<td>0.019(±0.001)</td>
</tr>
<tr>
<td>$r$ [cm$^{-1}$]</td>
<td>0.116(±0.004)</td>
<td>0.107(±0.004)</td>
</tr>
</tbody>
</table>

Table 3.1: Mean extinction ($\kappa$), absorption ($k$) and scattering ($r$) coefficients for the range of 400 to 700 nm, measured in a depth of 6 to 12 cm below the snow surface.

3.5.2 Spectral radiation in the snow in the range of 350 to 2500 nm

The advantage of the spectroradiometer over the photovoltaic detectors is the high spectral resolution. But the spectroradiometer has only one measuring device. Therefore it is not possible to carry out simultaneous observations at different levels. Furthermore, the errors in the determination of the exact measuring height are estimated to be 10% of the extent of the overlying snow cover. The irradiances on different levels in the snow are shown in Figure 3.8 and 3.9 as examples for the measurements made with the spectroradiometer. In the figures both axes are chosen logarithmic.

On day 205 the measuring heights were only in the upper 5.1 cm of the snow cover, on day 218 measurements were made down to 62 cm below the snow surface. The data for wavelengths above 2300 nm and for absolute irradiances smaller 0.0007 W/m$^2$/nm can
3.5. Solar radiation penetrating in the snow cover

Fig. 3.8: Spectra of down-welling radiative fluxes on different levels in the uppermost 5.1 cm of the snow cover measured with a spectroradiometer. The measuring time is 19.40 to 20.10 UTC on day 205.

Fig. 3.9: Spectra of down-welling radiative fluxes on different levels down to 62 cm below the snow surface measured with a spectroradiometer. The measuring time is 17.05 to 17.35 UTC on day 218.

be regarded as noise. The incoming energy at the surface in the range of 350 to 2500 nm for the series on day 218 was 522.7 W/m² (100%). At 3.5 cm depth the irradiance was 99.1 W/m² (19% of surface value), while only 3.5 W/m² was observed at 62 cm depth. The incoming energy for day 218 in the range of 400 to 700 nm was 244.2 W/m² (100%), in 3.5 cm depth 63.8 W/m² (26.1%) and in 62 cm 1.38 W/m² (0.6%). The data show a
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strongly varying spectral composition with depth, with a depletion mainly in the wavelengths longer than 1000 nm (see Figure 3.8 and 3.9). On day 205 the first measuring height is 0.8 cm below the surface, but smaller irradiances have been measured than on day 218 at a depth of 3.5 cm. The reason for this discrepancy must be, at least partly, in the determination of the exact measuring heights. Another potential cause is the difference in the properties of the uppermost snow layer. The uppermost snow layer plays a major role for the penetrating radiation.

3.5.2. a Extinction as a function of wavelength and depth

The extinction coefficients for the downward fluxes were determined by linear regressions for each wavelength. In Figure 3.10 a few examples are shown. In the top panel the regressions are made with all data in the snow but omitting the surface measurements while in the bottom panel all data including the surface measurement are considered.

A comparison of the plots in Figure 3.10 shows clearly that, if the surface measurement is included in the regression, the law of exponential extinction is not in good agreement with the experimental data. This is evidence that a great portion of energy is reflected and absorbed at the interface snow-atmosphere.

3.5.2. b Extinction as a function of the snow properties

Figure 3.11 shows mean values of the extinction coefficients for 400 to 700 nm with depth. In the same figure a profile of the mean snow density is shown. The values of the extinction coefficients for 400 to 700 nm are summarized in Table 3.2.

<table>
<thead>
<tr>
<th>day 172</th>
<th>day 205a</th>
<th>day 205b</th>
<th>day 218</th>
<th>day 223</th>
</tr>
</thead>
<tbody>
<tr>
<td>z [cm]</td>
<td>κ</td>
<td>z [cm]</td>
<td>κ</td>
<td>z [cm]</td>
</tr>
<tr>
<td>0.6</td>
<td>0.765</td>
<td>0.8</td>
<td>1.685</td>
<td>1.7</td>
</tr>
<tr>
<td>2.8</td>
<td>0.163</td>
<td>5.0</td>
<td>0.287</td>
<td>1.2</td>
</tr>
<tr>
<td>8.2</td>
<td>0.136</td>
<td>10.0</td>
<td>0.084</td>
<td>1.6</td>
</tr>
<tr>
<td>12.5</td>
<td>0.115</td>
<td>15.0</td>
<td>0.100</td>
<td>2.9</td>
</tr>
<tr>
<td>16.5</td>
<td>0.098</td>
<td>20.0</td>
<td>0.070</td>
<td>3.6</td>
</tr>
<tr>
<td>22.0</td>
<td>0.103</td>
<td>30.0</td>
<td>0.081</td>
<td>4.1</td>
</tr>
<tr>
<td>32.5</td>
<td>0.109</td>
<td>40.0</td>
<td>0.079</td>
<td>41.0</td>
</tr>
<tr>
<td>36.5</td>
<td>0.120</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3.2: Average mean values of the extinction coefficients between two levels in the range of 400 to 700 nm within the snow.

While the snow density increases until 45 cm depth, the extinction increases until 15 - 25 cm and then remains roughly constant between 0.05 and 0.15 cm\(^{-1}\) with increasing depth. The variation of the extinction, however, is not only a function of the snow density. Another characteristic that changes with increasing snow depth is the shape of the snow grains. In the upper layers the grains are small and have edges. Further down the snow grains undergo a constructive metamorphosis and become larger and round. This is also the reason for a decreasing snow density around 45 cm depth.
Fig. 3.10: Down-welling radiative fluxes (in logarithmic scale) in the snow cover on different levels for selected wavelengths. In the upper panel only measurements below the snow surface are used while in the panel at the bottom all measurements are used for the linear regressions.
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3.5.3 Discussion on radiation in the snow

3.5.3.a Integrated measurements, 400 to 700 nm

Extinction, absorption and scattering coefficients were calculated by applying the theory of radiation in the snow (Ohmura, 1980). The average extinction coefficient was 0.071 cm\(^{-1}\) for downward radiation and 0.065 cm\(^{-1}\) for upward radiation. Average values of \(k\) and \(r\) are shown in Table 3.1. The extinction in the considered bandwidth was found to be higher for downward radiation than for upward radiation. A hypothesis for this might be the selective absorption and the directional scattering properties of the snow or the self-shading.

Furthermore, a small dependence of the extinction on the zenith angle was found. But the data revealed opposing trends for downward and upward radiation (see Fig. 3.5). An explanation for the temporal variation of the extinction could be the variation of the spectral composition of the incoming solar radiation with time. King and Simpson (2001) found in their monochromatic investigations a weak dependence of the extinction on the solar zenith angle. 24% of the total incoming radiation in the range of 400 to 700 nm penetrate down to a depth of 6 cm and 16% to a depth of 12 cm. The reflectivity, \(r_s\), within the snow cover was found to be in the range of 0.5 to 0.6, slightly increasing with depth. The average total reflectivity of the snow cover, \(\alpha\), was 0.94(±0.01) in the range of 400 to 700 nm (see Figure 3.7).

3.5.3.b Spectral measurements, 350 to 2500 nm

Measurements of downward radiation fluxes on different levels were carried out with a spectroradiometer. Measuring heights were chosen very close to the snow surface down to a depth of 62 cm. Again the attenuation of the radiation was determined. Spectral extinction coefficients were calculated for different levels. The snow cover absorbed wavelengths...
longer than 1200 nm in the uppermost millimeters, wavelengths longer than 1000 nm in the uppermost 3 cm. Close to the surface extinction coefficients proved to be higher for all wavelengths, decreasing with increasing depth. Generally, wavelengths in the range of 400 to 550 nm yielded the smallest attenuation. The extinction increased rapidly above 550 nm. In general downward into the snow a decreasing extinction goes together with an increasing density. For wavelengths longer 1000 nm the absolute values of the irradiances are small and less relevant for energy balances. The calculated extinction coefficients for wavelengths longer 1000 nm displayed a wide scatter. For wavelengths in the range of 350 to 800 nm, the extinction coefficients vary from 0.8 cm\(^{-1}\) near the snow surface to 0.015 cm\(^{-1}\) at 62 cm depth. Only 10% or less of the total incoming radiation in the range of 350 to 2500 nm reach a depth of 5 cm. In 30 cm depth the downward irradiance is reduced to 1% of the surface irradiance. The percentage of radiation contained in the wavelengths of 400 to 700 nm increases rapidly with depth from 48% at the surface to 75% around 10 cm below the surface.

3.5.3.3 Comparison with other studies

On the west coast of Greenland Konzelmann (1994) made measurements on the average equilibrium line altitude. He collected data mostly in wet snow and reports only few results from dry snow observations. Konzelmann observed the lowest extinction coefficients in wet snow and the highest in dry snow. The values vary between 0.058(±0.004) (wet snow, 400–700 nm) and 0.098(±0.007) (dry snow, 400–700 nm). But the data have to be compared with caution. At Summit, measurements were made down to 12 cm, while Konzelmann measured at 10 cm, 20 cm and 40 cm depth. Also the physical properties of the snow like the grain size and grain shape differ at the two locations. Konzelmann’s values are in general smaller than the one’s measured at Summit. This can be explained by the difference in liquid water content. At Summit the snow is completely dry. Several studies mention lower extinction in wet snow (for example in Gerland et al. (2000), Konzelmann (1994) and Fukami et al. (1985)). Only few authors measured extinction in dry snow. In some studies one value is given for the extinction along with a specification of the density or grain size (for example in Konzelmann (1994), Sergent et al. (1987), Fukami et al. (1985) and Mellor (1966)). In others studies only trends of the extinction with varying wavelength and increasing depth are given (for example in Brandt and Warren (1993) and Weller and Schwertfeger (1970)). Olsson (1936) mentions a value for \(\kappa\) of 0.7 for dry snow which would be in good agreement with this study, but he measures in a wider range. Weller and Schwertfeger (1970) find \(\kappa = 0.110\) near the surface and \(\kappa = 0.022\) in 1 m depth. This is in good agreement with the spectroradiometer measurements of the present work if an integrated value over all wavelengths is considered. Liljequist (1956) data show similar trends compared to this study. His measurements were made in 20 to 100 cm depth, but he reports only one value for the extinction for all depths. Gerland et al. (2000) found a constant extinction coefficient in the range of the photosynthetically active radiation (400–700 nm) which is a good approximation, as shown in the present study.
3. Basic Concepts of Snow Radiative Processes

3.5.3.d Conclusion

In general, the comparability of different studies is difficult due to high variability of snow properties and the different instruments. But there is general agreement in certain points. Experimental data show, with only a few exceptions, decreasing attenuation with increasing density. Wavelengths longer than 1000 nm are absorbed very near the surface. The snow acts as a blue color filter. This also becomes apparent from the blue light in snow caves.

King and Simpson (2001) state that in the near-infrared spectral region, snow has relatively high absorption strength compared to its scattering strength. In contrast, in the visible and UV regions, the absorption of light by snow is weak, and scattering becomes dominant. The depth to which light penetrates into the snow pack is a combination of the scattering and absorption strength. A light beam is attenuated more effectively, due to absorption, penetrating a large snow grain than a small one. But the attenuation due to scattering in fine-grained snow is much higher than in a snow layer of the same thickness with coarse-grained snow due to an increased total surface area of the snow grains (Brandt and Warren, 1993). Several authors made models for the radiation transfer in the snow. Glendinning and Morris (1999) present an incorporation of spectral and directional radiative transfer in a snow model. This model takes into account that scattering of light in snow is highly wavelength-dependent and that snow grains scatter solar radiation anisotropically with significant forward and multiple scattering. Gerland et al. (2000) describe a physically-based two-stream one-dimensional model including the spectral dependence of solar radiation penetrating snow and an extinction coefficient variable with depth. Peterson et al. (2002) present a one-dimensional Monte Carlo model of photon scattering within a snow pack. Gerland et al. (2000) and Peterson et al. (2002) validate their models with experimental data from one site only and find good agreement with the models.

To simulate radiation in the snow accurately for near surface snow, radiative transfer models must include some spectral details because the penetration depth of solar radiation varies across the solar spectrum. Radiation in the visible range can be measured at more than 50 cm depth while radiation with wavelengths longer than 1500 nm is absorbed in the top few millimeters. By using 'average' values for penetration depth and extinction coefficients, absorbed radiation in the snow is highly overestimated (Brandt and Warren, 1993).

3.5.3.e Further investigations

Accurate description of radiation in snow is dependent on spectral details of incoming radiation and on the spectral behavior of the flux parameters (extinction, absorption and scattering coefficients). However, there is a strong variation of the flux parameters with changing snow properties. Experimental validation of snow radiative transfer models, taking spectral characteristics and physical snow properties into account is necessary. Furthermore, the surface absorptivity needs to be investigated to get a better understanding of the processes at the interface snow-atmosphere and the resulting effects on the energy balance components of a snow cover.
Chapter 4

IACETH Goniospectrometer: A Tool for Hyperspectral HDRF Measurements

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Abstract

This work presents a new field goniospectrometer developed at the Institute for Atmospheric and Climate Science (IAC) of the Swiss Federal Institute of Technology (ETH; Switzerland). The goniospectrometer was built to study the Hemispherical Directional Reflectance Factor (HDRF) of snow, but can also be applied to other surfaces with moderate surface roughness.

The IACETH goniospectrometer measures HDRFs with high spatial resolution. The goniometer is exclusively built of straight parts, thus ensuring a high pointing accuracy. The two robotic arms are controlled automatically with step motors, whereby the step size can be defined by the user. With the default grid size of $15^\circ$ in zenith and azimuth, the time needed to collect one complete HDRF dataset is 11 minutes, corresponding to a change of less than $4^\circ$ in solar zenith and azimuth angles.

The spectrometer comprises two probes. The first probe is equipped with a $3^\circ$ foreoptic and is used for taking a spectrum of the reflected radiance; the second is placed on a tripod, has a $2\pi$-foreoptic and simultaneously records a spectrum of the incoming irradiance. Both probes measure in the spectral range from 350 to 1050 nm, with a resolution of approximately 3 nm at around 700 nm.

The performance of the new goniospectrometer was tested at the Greenland Environmental Observatory Summit Station ($72^\circ$ 35’ N, 34$^\circ$ 30’ W, 3203 m ASL) during the summer of 2004.
4. **IACETH Goniospectrometer: A Tool for Hyperspectral HDRF Measurements**

### 4.1 Introduction

The surface albedo is of great importance for understanding the energy balance of the earth's surface. Satellites provide an extraordinary means to measure reflected sunlight, especially in remote regions. However, the measured reflectances need to undergo complex data processing to account for the illumination conditions and the view angles of the airborne or space-based sensors relative to the target. The factor describing the angular distribution of reflected radiance is called Hemispherical Directional Reflectance Factor (HDRF), where hemispherical incoming irradiance and directional outgoing radiance is considered. The HDRF is defined as the ratio of the radiance outgoing from a surface into a specific direction to the radiance outgoing from a Lambertian surface under the same hemispherically-integrated incoming irradiance. The HDRF is a function of four angles: incoming (solar) zenith and azimuth ($\theta_i, \varphi_i$), and outgoing (reflected) zenith and azimuth ($\theta_r, \varphi_r$). Furthermore, the HDRF strongly depends on the physical surface properties (e.g. Warren, 1982; Steffen, 1987; Martonchik, 1994; Warren et al., 1998; Pirazzini, 2004; Kokhanovsky and Zege, 2004).

With the advent of new satellite systems that offer hyperspectral resolution and off-nadir tilting capabilities, there is an increasing need for multidirectional and hyperspectral ground truth data (Sandmeier et al., 1998; Sandmeier, 2000; Liang et al., 2000; Nolin and Liang, 2000). In situ measurements of HDRF datasets consist of a combination of multidirectional and hyperspectral data, collected within a time period that is as short as possible to make changes in the solar illumination geometry negligible. As a consequence of these demands the existing database of experimental HDRF measurements is still small. Following are a few automated field goniometric instruments that have been utilized for snow HDRF measurements:

- a Portable Apparatus for Rapid Acquisition of Bidirectional Observations of the Land and Atmosphere (PARABOLA; Deering and Leone (1986); Abdou et al. (2001)),
- a Field Goniometer System (FIGOS; Sandmeier and Itten (1999)), and
- an Automated Spectro-Goniometer (ASG; Painter et al. (2003)).

Complete instrumentation for HDRF measurements consists of a goniometer placing a radiometric sensor in the desired position on the hemisphere. While PARABOLA measures the radiance in eight spectral bands, FIGOS and ASG utilize high resolution spectrometers similar to the one used for the measurements described in this study.

This work describes a field goniospectrometer that is capable of measuring HDRFs of surfaces with moderate surface roughnesses like snow, sand, soil or pavement. The instrument was developed at the Institute for Atmospheric and Climate Science at the Swiss Federal Institute of Technology (IACETH). The main motivation was to study the HDRF of dry snow on the Greenland ice sheet and to investigate its influence on the surface energy balance and on the hemispherical albedo.

The IACETH field goniospectrometer measures the HDRF with high spectral and spatial resolution. It is fully automatic and the measuring program can easily be adapted in the field, allowing to select the appropriate sampling resolution for the HDRF measurements. The new instrument was designed for use in remote places which gave rise to a construction that is easily transportable and set up, but nevertheless is very robust and adjustable. It was tested and utilized at the Greenland Environmental Observatory Summit Station (72° 35’ N, 34° 30’ W, 3203 m ASL).

The emphasis of this work is on the technical description of the IACETH goniospec-
trometer and its functionality. Necessary calibrations and data processing routines, including the investigation of measurements affected by self-shadowing, are also discussed.

A short theoretical overview on directional reflectance and the commonly used equations are given in section 4.2. Section 4.3 describes the hardware components of the new system, consisting of the IACETH goniometer and a portable spectrometer, Fieldspec Pro Dual VNIR spectrometer manufactured at Analytical Spectral Devices (ASD), in Boulder, Colorado. It also presents the utilized Spectralon reference panel along with the corrections needed to account for its departure from the ideal Lambertian reflector. Details of the hardware control, the data acquisition, and the data processing routines, as well a brief discussion of self-shadowing and an evaluation of the pointing accuracy of the goniometer, are also included in this section. Section 4.4 follows with first results, and a short discussion and some conclusions are provided in section 4.5.

4.2 Theoretical background

The directional reflectance properties of a surface can be described by a set of functions. Complete definitions of these functions are given in Nicodemus et al. (1977) and Martonchik et al. (2000). Relevant for the present work are the Hemispherical Directional Reflectance Factor (HDRF) and the hemispherical reflectance, $\rho$ (albedo).

It is well known that snow reflects solar radiation anisotropically and that snow surfaces are far away from being Lambertian reflectors (e.g. Bohren and Barkstrom, 1974; Choudhury and Chang, 1981; Warren, 1982; Steffen, 1987). The HDRF is thus defined as the ratio of the radiance reflected by a surface in a specific direction to that reflected in the same direction by a perfect Lambertian surface under ambient illumination (Martonchik et al., 2000). Assuming that the diffuse component of the incoming irradiance is isotropic, the HDRF can be expressed as

$$\text{HDRF}_\lambda(\theta_r, \varphi_r, \theta_i, \varphi_i) = \frac{L_\lambda(\theta_r, \varphi_r)}{L_{\lambda,Lamb}} = \frac{L_\lambda(\theta_r, \varphi_r)}{\left(\mu_i E_{\lambda,dir}(\theta_i, \varphi_i) + E_{\lambda,diff} \pi\right)} , \quad (4.1)$$

where $L_\lambda$ denotes the spectral radiance; $L_{\lambda,Lamb}$ the radiance of a Lambertian surface; $E_{\lambda,dir}$ the direct incoming irradiance; $E_{\lambda,diff}$ the diffuse incoming irradiance; $\theta_r$ and $\varphi_r$ the reflection zenith angle and azimuth angle; $\theta_i$ and $\varphi_i$ the incident zenith angle and azimuth angle; and $\mu_i$ the direction cosine of the incident solar beam.

With the appropriate instrumentation, HDRFs can be measured directly in the field. It is common to utilize a white reference standard, for instance a Spectralon panel (see section 4.3.3), as a Lambertian surface. However, commercially available panels are not perfect. Departures from a perfect reference are considered by introducing a spectral correction term $C_{\lambda,corr}$ that accounts for the directional reflectance characteristics and for subunity in total reflectance:

$$\text{HDRF}_\lambda(\theta_r, \varphi_r, \theta_i, \varphi_i) = \frac{L_\lambda(\theta_r, \varphi_r)}{L_{\lambda,Spectralon}(\theta_i=0, \varphi_i=0)} \frac{1}{C_{\lambda,corr}} \quad . \quad (4.2)$$

As seen in Eq. (4.1) the HDRF is a function of the wavelength and four angles (Fig. 4.1). In experiments, HDRF datasets are measured under arbitrary illumination geometries. For easier comparability, however, the results are usually presented with respect to the
Fig. 4.1: Angles and naming conventions for defining the HDRF: $E_{i,\text{dir}}$ = direct incoming irradiance, $E_{i,\text{diff}}$ = diffuse incoming irradiance, $L_r$ = reflected radiance, $\theta_i$ = Solar zenith angle, $\varphi_i$ = Solar azimuth angle, $\theta_r$ = reflection zenith angle, and $\varphi_r$ = reflection azimuth angle.

solar principal plane (e.g. Steffen, 1997; Sandmeier, 2000; Painter and Dozier, 2004b). The solar principal plane is defined in such a way that the azimuth of the incident radiation is equal to 180° and the forward scattering direction is equal to 0°. This reduces the number of relevant angles to three: $\theta_i, \theta_r$ and $\varphi_i - \varphi_r$.

In some earlier studies, researchers used the expression bidirectional reflectance factor, (BRF; theoretically considering incoming and outgoing directional radiances) when talking about the HDRF. Obtaining the BRF through ground measurements is only possible by measuring the HDRF and applying corrections for the diffuse part of incoming radiance (Sandmeier, 2000; Bruegge et al., 2001).

By integration of the HDRF$_\lambda$ over the hemisphere the spectral albedo $\rho_\lambda$ can be determined as follows:

$$\rho_\lambda(\theta_i, \varphi_i) = \frac{1}{\pi} \int_0^{2\pi} \int_0^{\pi/2} \text{HDRF}_\lambda(\theta_i, \varphi_i, \theta_r, \varphi_r) \cos(\theta_r) \sin(\theta_r) d\theta_r d\varphi_r ,$$

and by the integration over all wavelengths of the spectral albedo weighted by the spectral irradiance $E_\lambda$, the broadband albedo $\rho$ can be calculated.
4.3 Instrument description

4.3.1 IACETH goniometer

The IACETH goniometer was designed for measuring the HDRF of relatively smooth surfaces. With this in mind, the radius of the measuring hemisphere was set to 1 m, defining a distance of 1 m between sensor and target. To avoid disturbances of the surface the sensor-arms were attached to an extended boom rather than placed in the middle of a large construction set on the surface, which is the case for most of the previous field goniometers.

Analysis of the motions necessary to let a point sweep a hemisphere – partly done with the aid of a CAD system – suggested that bent parts, as used on existing goniometers (FIGOS and ASG), are not needed. We thus use two straight arms of equal length, but different cross sections (the arms are reduced in mass by milling off statically unnecessary material). The first is attached under an angle of 45° to a horizontal boom and rotates about a vertical axis z. It carries the second arm, which folds back a full 180° about an inclined axis k (Fig. ??). Rotating these two arms individually by \( \varphi_r \) (motor I) and \( \eta \) (motor II) enables the sensor to be positioned on any point of the hemisphere above the target. The first angle is the same as the view azimuth of Eq. (4.1). The relation between \( \eta \) and the view reflectance angle \( \theta_r \) can be derived either from Rodrigues’ formula (Tsai, 1999) or using spherical trigonometry. It reads,

\[
\cos \eta = \frac{\cos \theta_r - \cos a \cos b}{\sin a \sin b}, \tag{4.4}
\]

where \( a \) and \( b \) are the angles subtended by the two moving arms. With \( a = b = \pi/4 \),
given by the instrumental geometry, one obtains:

\[
\cos \eta = \frac{\cos \theta_r - \frac{1}{2}}{\frac{1}{2}} = 2 \cos \theta_r - 1.
\]  

(4.5)

Values of \( \eta \) obtained from Eq. (4.5) for the default values of \( \theta_r \) are listed in Table 4.1.

To move the two arms we chose stepping motors of an 'intelligent' type with integrated drive electronics. A gearbox was introduced as well to achieve the necessary torque. The possibility to adjust the motor’s driving and halt forces individually made it unnecessary to use a clutch. First tests showed a rather small backlash of the gearbox that made the arm assembly oscillate for a short while after the top drive controlling the azimuthal position stopped. This problem was removed by introducing a disc brake for the top drive. Technical details of the motors are listed in Table 4.2.

<table>
<thead>
<tr>
<th>( \theta_r )</th>
<th>15°</th>
<th>30°</th>
<th>45°</th>
<th>60°</th>
<th>75°</th>
<th>80°</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \eta )</td>
<td>21.3°</td>
<td>42.9°</td>
<td>65.5°</td>
<td>90.0°</td>
<td>118.8°</td>
<td>130.8°</td>
</tr>
</tbody>
</table>

Table 4.1: Values of \( \eta \) (rotation about z-axis) obtained from Eq. (4.5) for the default view zenith angles \( \theta_r \)

<table>
<thead>
<tr>
<th></th>
<th>Top drive (Azimuth)</th>
<th>Lower drive (Zenith)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type</td>
<td>MDMF 1719-G-51:1-F</td>
<td>MDDBC 1410-G-51:1-F</td>
</tr>
<tr>
<td>Step Frequency max</td>
<td>2 MHz</td>
<td>1 MHz</td>
</tr>
<tr>
<td>Counts per revolution</td>
<td>20356</td>
<td>20356</td>
</tr>
<tr>
<td>Holding Torque max</td>
<td>53 Ncm</td>
<td>7 Ncm</td>
</tr>
<tr>
<td>Detent Torque max</td>
<td>2.8 Ncm</td>
<td>1.0 Ncm</td>
</tr>
<tr>
<td>Weight</td>
<td>335 g</td>
<td>142 g</td>
</tr>
<tr>
<td>Gearbox:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gearbox Ratio</td>
<td>51:1</td>
<td>51:1</td>
</tr>
<tr>
<td>Output Torque max</td>
<td>15 Nm</td>
<td>4.5 Nm</td>
</tr>
<tr>
<td>Weight</td>
<td>610 g</td>
<td>264 g</td>
</tr>
</tbody>
</table>

Table 4.2: Specifications for the drives (Intelligent Motion Systems)

The stepping motors offer a total of 20356 counts per revolution. Obviously, not all steps in \( \varphi_r \) or \( \eta \) can be realized exactly with an integer number of counts. However, the rounding error is at most of one count. This corresponds to a angular accuracy of roughly 1°.

As seen in Fig. 4.3, the whole assembly consists of a vertical post, a horizontal boom fixed on that post and two moving arms attached to the end of the boom. The horizontal boom of the goniometer can be rotated completely around the main vertical post. However, for the experiments at the Summit Station of the Greenland Environmental Observatory the boom was always aligned north-south, with the moving arms holding the radiometric sensor attached in the south. This guaranteed that consecutive experiments viewed the same surface area. The field deployment is shown in Fig. 4.4.

To ensure that the sensor is viewing the same target point on the ground at any time and from every position, (a) the instrument has to be level and (b) the sensor has to have
4.3. Instrument description

Fig. 4.3: Design of the IACETH Goniospectrometer

an exact distance of one meter from the target area in every position. Therefore, (a) a three point leveling plate is mounted on the vertical post (Fig. 4.3) and (b) the vertical post can be extended with a fine-threaded screw located inside the vertical post. The distance to the target point on the surface is measured with a laser distance meter.

4.3.2 Spectrometer

The IACETH goniometer is operated with a standard field spectrometer, FieldSpec Pro Dual VNIR, from ASD (see additional information online at http://www.asdi.com). This spectrometer comes with two 1.5-m fiberoptics, which were replaced by two 4-m optics given the size of the goniometer. The two 4-m optics were tested at ASD and were shown to have the same properties as the 1.5 m optics. The advantage of having two probes is that measurements of the incoming spectral irradiance can be taken simultaneously with those of spectral reflectance with the same instrument.

The utilized spectrometer covers a nominal spectrum between 350 and 1050 nm. The spectrum is measured with a 512-channel silicon photodiode array. Each channel, an individual detector itself, is geometrically positioned to receive light within a narrow bandwidth (1.4 nm). The VNIR spectrometer has a spectral resolution [full width half-maximum (FWHM) of a single emission line] of approximately 3 nm at around 700 nm.

For our measurements we attached a $3^\circ$ field-of-view foreoptic to the fiber measuring
the reflected radiance. A $2\pi$ foreoptic was mounted on the reference fiber measuring the incoming irradiance. While the target fiber is fixed along the goniometer arm, the reference foreoptic is put on a tripod and placed beside the goniometer (Fig. 4.4).

Accurate determination of absolute radiometric values depends to the greatest extent upon accurate calibration of the utilized spectrometer. Both the $3^\circ$ radiance and $2\pi$ irradiance foreoptic were calibrated radiometrically at ASD just before the instrument was deployed in the field. Information on the calibration procedure is available on the web site of ASD (online at http://www.asdi.com).

### 4.3.3 Spectralon panel

According to Eq. (4.1), the HDRF is defined with respect to a Lambertian reference. In our case, a near-Lambertian Spectralon panel from Labsphere (information available online at www.labsphere.com) was utilized. The Spectralon has been calibrated at the factory with an illumination angle of $8^\circ$. For each wavelength a calibration coefficient is provided that accounts for the subunity in reflectance of the panel. The values of this coefficient lie between 0.982 and 0.988 for wavelengths in the range of 350 - 1050 nm. However, as shown by Sandmeier et al. (1998), even a calibrated Spectralon can show deviations of up to 10% from the perfect Lambertian behavior, depending on the angle of the incident beam. For this reason and given the fact that the anisotropy is nearly invariant for all Spectralon panels (Bruegge et al., 2001) we apply the correction algorithm of Sandmeier et al. (1998) to our data.
4.3.4 Control of the hardware, data acquisition

The motion of the goniometer is controlled with a Campbell CR10X datalogger. The datalogger is individually connected to the two motors of the goniometer and to the laptop, which controls the spectrometer. A default program for a standard sampling routine was stored on the datalogger. The program, however, can easily be modified in the field, in particular if a higher sampling resolution is needed.

A complete cycle starts with an optimization of the sensor on the actual reflectance of the Spectralon panel, followed by a nadir reflectance measurement on the Spectralon. The Spectralon panel is fixed on a turnable boom (Fig. 4.3). After turning the Spectralon panel away, a spectrum of the reflected radiance at nadir is taken. Then, motor II (Fig. 4.3) moves the lower arm in a new viewing zenith position. In this position the upper arm is turned counterclockwise around the vertical axis providing a scan of all azimuth positions. After this first cycle, the lower arm moves to a new zenith position and the azimuth is now scanned clockwise. At the end of a complete hemispherical cycle the two arms are moved back into their home position and a second snow nadir and Spectralon reflectance sample are saved. Sometimes one of the measurements is outside the optimization range defined at the beginning of each cycle. In this case, the program controlling the movement can be interrupted and a new optimization can be carried out.

Fig. 4.5 shows the measuring path of a complete sampling cycle on a polar plot for the chosen sampling resolution of 15°. The time needed to sample one point is 5 s on average, including the traveling time to a new position. For a resolution of 15° the time required to complete a cycle is 11 minutes. At the location of the Summit Station at the...
Greenland Environmental Observatory, this corresponds to a change of less than 4° in the solar azimuth angle.

Each time the sensor arrives at a measuring position, a serial command is sent to the laptop controlling the spectrometer and a spectrum is saved. Spectra of the reflected radiance and incident irradiance are saved individually. The files are named automatically with consecutive numbers. A complete cycle results in a total of 296 files, each containing 701 values for the wavelength range between 350 and 1050 nm.

In summary, for a sampling resolution of 15° a full HDRF cycle consists of 148 measured positions, including 144 measurements distributed over the hemisphere (6 zenith angle positions and 24 azimuth angle positions), two surface nadir measurements, and two measurements on the Spectralon panel (nadir)– one at the beginning and one at the end of each cycle.

4.3.5 Pointing accuracy

The pointing accuracy of the goniometer was investigated with the aid of a laser pointer. The pointer was attached to the head of the lower arm (replacing the radiometric sensor), and a complete sampling cycle was carried out, recording the path left by the laser beam on the surface (Fig. 4.6). For the ideal distance of 1 m between laser pointer and target, the area covered by the laser beam was of 1.4 cm in diameter, implying a pointing accuracy of roughly ±1 cm. Since the 3° field-of-view foreoptic results in a elliptical footprint with a major diameter of 5.2 cm at nadir and 31 cm at 80° (Fig. 4.7), this accuracy is considered as satisfactory.

However, the pointing accuracy is very sensitive with respect to the distance. Our experiment revealed that changing the height of the horizontal boom by ± 1 cm degrades the pointing accuracy by an order of magnitude, with the result that the laser beam spans an area of 15 cm in diameter at $\theta_r = 80°$. 

\[ \text{Fig. 4.6: Results of the pointing accuracy experiment; 'M' denotes the actual target when the arm is at the 'home' position ($\theta_r = 0°, \varphi_r = 0°$). During a complete cycle the laser beam traveled within the shaded area.} \]
4.3.6 Self-shadowing

It is unavoidable that a shadow caused by the moving arm and the foreoptic of the instrument crosses the target area several times during a complete cycle. Occultation occurs when parts of the goniometer align with the sun and the target center. Examples of self-shadowing are shown in Fig. 4.8. These data were collected on 26 June, 0836 and 1236 (LT) at solar zenith angles of 58° and 49°. In these examples, and for all other data collected during the 2004 field campaign the measurements affected by self-shadowing are always in the 90° - 180° quadrant after projection into the solar principal plane. Data points with low HDRFs values due to self shadowing are clearly recognizable in Fig. 4.8. In the course of the data processing, these shadowed data points are filtered and replaced with corresponding values from the other side of the solar principal plane, assuming symmetry along the solar principal plane. In general, knowledge of the instrumental geometry and the position of the sun is sufficient for determining the area affected by self-shadowing.
4.4 Performance of the IACETH goniospectrometer in the field

4.4.1 Experimental site

An extended set of HDRF data was collected in the summer season of 2004 on the Greenland ice sheet, at the Greenland Environmental Observatory Summit Station (72° 35’ N, 34° 30’ W, 3203 m ASL) (Ohmura, 2001). Complete hemispherical cycles were measured on clear sky days every one or two hours for a total of 90 hemispheres. The largest solar zenith angle was 77° and the smallest 49°. It was pointed out in the introduction that, in addition to the illumination geometry, the physical properties of the snow surface are im-
portant parameters influencing the HDRF of snow. Although the experimental site lies in the dry snow zone where no melting occurs, the snow surface properties vary significantly within a short time. During the summer 2004, the relevant snow characteristics were measured in the field. These observations indicate that the occurring snow surfaces can be grouped in four classes: new snow with unbroken hexagonal snow crystals, wind-broken small snow grains, rounded snow grains from snow metamorphism and coarse surface hoar growing during riming events.

4.4.2 Results

Results of the experiments are presented in Figs. 4.9 - 4.11. The datasets are plotted using the same format as in Fig. 4.8. For the data analysis and visualization the Interactive Data Language (IDL; information available online at http://www.rsinc.com/idl/) was utilized.

The HDRF shows a large variation across the solar spectrum. Fig. 4.9 shows HDRF datasets, measured on 1835 LT 26 June 2004 at a solar zenith angle, $\theta_i$, of 67°. The target area was covered with large crystals of surface hoar. Three wavelengths (400 nm, 550 nm and 1000 nm) have been chosen to depict the variation of the reflectance with wavelength. The degree of anisotropy of the distribution of the HDRF is given by the Anisotropy Index (Anix), defined as the ratio of the maximum and the minimum HDRF over the hemisphere for a given wavelength,

$$\text{Anix}_\lambda = \frac{\max(\text{HDRF}_\lambda)}{\min(\text{HDRF}_\lambda)}.$$  

(4.6)

The Anix values for Fig. 4.9 are 1.75 (400 nm), 1.58 (550 nm) and 2.72 (1000 nm).

The HDRF also shows a strong dependence on the illumination conditions. Fig. 4.10 shows the HDRF for four different solar illumination geometries for the wavelength of 550 nm. The datasets were collected in 2-h time intervals on 26 June 2004, starting at solar noon. The surface was again covered with hoar. The values of the Anix are 1.40 ($\theta_i = 49^\circ$), 1.48 ($\theta_i = 51^\circ$), 1.62 ($\theta_i = 58^\circ$) and 1.59 ($\theta_i = 67^\circ$).

Fig. 4.11 shows the variation of the HDRF with changing snow properties. The two panels on the left show data collected on a snow surface of small windbroken crystals, whereas the two panels on the right stem from a snow surface covered with large crystals from surface hoar. As an example, two wavelengths, 550 and 1000 nm, are displayed. Both datasets are collected at a solar zenith $\theta_i$ of 51°. For small crystals the value of the Anix is 1.42 (550 nm) and 1.60 (1000 nm), and for large crystals 1.13 (550 nm) and 1.28 (1000 nm).

In general, our preliminary results confirm findings from previous experiments. A strong forward-scattering peak is visible in the majority of all HDRF datasets (e.g., Steffen, 1997; Aoki et al., 2000; Painter, 2002). A weak backward-scattering peak is revealed at special conditions, for example, large illumination and measuring zenith angles. The Anix increases with increasing solar zenith angle and shows high values for surfaces covered with new snow.

4.5 Summary

A new field goniospectrometer is presented, including its technical details and functionalities. The hardware components, the control mechanisms, and the data acquisition
The motion of the IACETH goniometer is controlled with a standard datalogger. A program was written to guide the radiometric sensor automatically through a complete cycle. With a default sampling resolution of 15° in zenith and azimuth the program takes 11 minutes to collect a full HDRF dataset. The time needed for a complete cycle corresponds to a change of less than 4° in the solar zenith and azimuth. The spatial

Fig. 4.9: Polar plots of the HDRF for all view angles and 3 different wavelengths, measured on 26 June 2004. The solar zenith angle $\theta_i$ was 67°. The values for the Anix are 1.75 (400 nm), 1.58 (550 nm) and 2.72 (1000 nm), respectively. The data are shown in the same format as in Fig. 4.8
Fig. 4.10: Polar plots of the HDRF for all view angles and four different illumination conditions. The collection time is 1236 (solar noon), 1437, 1636 and 1835 LT 26 June 2004, with solar zenith angles of 49°, 51°, 58° and 67°. The wavelength is 550 nm. The values of the Anix are 1.40 (θ<sub>i</sub> = 49°), 1.48 (θ<sub>i</sub> = 51°), 1.62 (θ<sub>i</sub> = 58°) and 1.59 (θ<sub>i</sub> = 67°). The data are shown in the same format as in Fig. 4.8.

resolution of the samples can be increased but at the cost of a prolonged assimilation time for completing a cycle. On average, the time needed for the collection of one data point is 5 s, including the traveling time to a new position.

The angular accuracy of the IACETH goniometer is very high, which results in a pointing accuracy of the sensor in the order of ±1 cm. The footprint size of the 3° foreoptic varies between 5.2 cm and 31 cm in diameter, depending on the view zenith.
4. IACETH Goniospectrometer: A Tool for Hyperspectral HDRF Measurements

Fig. 4.11: Polar plots of the HDRF collected on 2 days, where (left) the snow crystals on the surface were wind-broken and small (3 July 2004), and (right) the snow surface was covered with large crystals from surface hoar (27 June 2004). Both samples are taken at a solar zenith angle of 51°. The two upper plots show the HDRF for the wavelength of 550 nm and the lower two plots for 1000 nm. The values of the Anix for small crystals are 1.42 (550 nm) and 1.60 (1000 nm) and for the large crystals 1.13 (550 nm) and 1.28 (1000 nm). The data are shown in the same format as in Fig. 4.8.

With the spectrometer used in this study it was possible to measure incoming and reflected spectral radiance simultaneously. Having a continuous spectral reference of the incoming irradiance will be of advantage for the further analysis of the field data, allowing to calculate the HDRF with two different approaches.

The construction of the IACETH goniospectrometer is very robust and lightweight, which allows the instrument to be deployed in remote regions and to be left outdoors for several weeks.

Acknowledgment

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Chapter 5

A Field Study of the Hemispherical Directional Reflectance Factor and Spectral Albedo of Dry Snow

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Abstract

Hemispherical directional reflectance factors (HDRF) were collected under solar zenith angles from 49° to 85°. The experimental site was the Greenland Summit Environmental Observatory (72° 35’ N, 34° 30’ W, 3203 m above sea level) where both the snow and the atmosphere are very clean. The observations were carried out for two prevailing snow surface types: a smooth surface with wind-broken small snow grains and a surface covered with rime causing a higher surface roughness. A specially designed Gonio-Spectrometer (wavelength range 350 - 1050 nm) was developed at the Institute for Atmospheric and Climate Science and used to collect spectral HDRFs over the hemisphere. The angular step size was 15° in zenith and azimuth. The HDRFs showed strong variations ranging from 0.6 to 13, depending on the solar zenith angle. The HDRF distribution was nearly isotropic at noon. It varied with increasing solar zenith angle, resulting in a strong forward scattering peak. Smooth surfaces exhibited stronger forward scattering than surfaces covered with rime. At a solar zenith of 85°, an HDRF of ~13 was observed in the forward scattering direction for λ=900 nm. Spectral albedos were calculated by interpolating the HDRF data sets on a 2° grid and integrating individual wavelengths. Spectral albedos showed variations depending on the solar illumination geometry and the snow surface properties. Broadband albedos were calculated by integration of the spectral albedos over all wavelengths. The broadband albedos derived from directional measurements reproduced the diurnal pattern measured with two back-to-back broadband pyranometers.
5.1 Introduction

Surface albedo is the hemispheric reflectance integrated over the solar spectrum. It is a property of the surface and determines the disposal of solar energy at the earth’s surface which makes it a primary driver in the climate system. Snow and ice surfaces in particular exhibit a very high albedo, and a small change in albedo has a significant impact on their thermodynamics. Eventually, long-term changes in the energy balance of snow and ice lead to changes in thickness and extent, which feed back on the climate system with important consequences for the global temperature. Polar regions thus are expected to have the highest impact on global warming due to the albedo feedback (Hansen and Nazarenko, 2004; Alley et al., 2005).

Today’s state-of-the-art satellite sensors offer a method to measure and monitor polar snow and ice properties. For example visible- and near-infrared radiometers like MODIS (MODerate resolution Imaging Spectroradiometer) can be used to derive surface albedo on a high spatial and temporal resolution (Schaaf et al., 2002). However, the sensors on satellites subtend a certain angle to the surface target. This has to be taken into account especially when a surface is reflecting anisotropically like snow (Jin and Simpson, 1999, 2000; Painter and Dozier, 2004a). This may lead to under- or overestimation of the albedo derived from satellite measurements. Furthermore, most satellites measure in discrete spectral bands, making a narrow to broadband conversion algorithm necessary to obtain the value of the total reflected energy (e.g. Liang, 2001; Klein and Stroeve, 2002; Greuell et al., 2002; Greuell and Oerlemans, 2004).

Ground observations show that the reflectance of snow varies with snow type and snow surface roughness, wavelength, viewing angle and solar illumination conditions (Middleton and Mungall, 1952; Bohren and Barkstrom, 1974; Choudhury and Chang, 1981a; Wiscombe and Warren, 1980; Steffen, 1987). All these conditions are expressed with the Bidirectional Reflectance Distribution Function (BRDF) (Nicodemus et al., 1977; Martonchik et al., 2000; Schaepman et al., 2006), where both the illumination and reflectance is considered directional and only from a certain solid angle for each point on the hemisphere. A non-dimensional equivalent can be measured experimentally at the earth surface and is called the Hemispherical Directional Reflectance Factor (HDRF), where hemispherical incoming irradiance and directional outgoing radiance is considered. The HDRF is likewise a function of the surface characteristics and exhibits similar variations as the BRDF. The HDRF can best be parameterized from high quality ground observations. The BRDF can be derived from the HDRF only if the atmospheric composition is known to correct for the diffuse radiation which is included in HDRF measurement.

To the present day few studies have experimentally investigated the directional reflectance characteristics of snow (Steffen, 1997; Warren et al., 1998; Aoki et al., 2000; Li and Zhou, 2004; Painter and Dozier, 2004b; Bourgeois et al., 2006b). Accurate measurements of the anisotropic reflectance of snow with high spectral, angular and temporal resolution are rare. Comprehensive datasets are presented by Aoki et al. (2000), Li and Zhou (2004) and Painter and Dozier (2004b). Most of these experimental datasets only cover a limited range of either snow surface characteristics or illumination conditions. The advent of new satellite systems with multiangle measurements calls for more surface-based data (Jin et al., 2002; Klein and Stroeve, 2002; Tanikawa et al., 2002). A general HDRF parametrization for snow surfaces is needed.

This study presents measurements conducted at the Greenland Summit Environmental
5.2 Definitions and theoretical background

Observatory, 72° 35’ N, 34° 30’ W, 3203 m a.s.l., (Ohmura, 2001) in the dry snow zone on the ice sheet of Greenland. The instrumentation consisted of a specially designed device, the IACETH Gonio-Spectrometer (Bourgeois et al., 2006b) capable of measuring directional reflected radiation with high angular and spectral resolution. The collected HDRF datasets were extensively analyzed with respect to the hemispherical anisotropy, the illumination geometry and the snow surface characteristics. The spectral albedo of the various snow surfaces was evaluated with the same datasets, applying a spherical Delaunay interpolation and an integration. Finally the broadband albedo for the wavelength range 400 - 1000 nm was calculated and compared to the broadband albedo measured with two pyranometers for the spectral range 280 - 4000 nm.

5.2 Definitions and theoretical background

Only the relevant equations for this study, describing the anisotropic reflectance characteristics of a surface, are presented here. The Hemispherical Directional Reflectance Factor (HDRF) is defined as follows:

$$\text{HDRF}_\lambda(\theta_r, \varphi_r, \theta_i, \varphi_i) = \frac{L_\lambda(\theta_r, \varphi_r)}{L_{\lambda, \text{Lamb}}} = \frac{L_\lambda(\theta_r, \varphi_r)}{\left(\mu_l E_{\lambda, \text{dir}}(\theta_i, \varphi_i) + E_{\lambda, \text{diff}}\right)^{-1}}, \quad (5.1)$$

where $L_\lambda$ denotes the spectral radiance, $L_{\lambda, \text{Lamb}}$ the radiance of a Lambertian surface, $E_{\lambda, \text{dir}}$ the direct incoming irradiance, $E_{\lambda, \text{diff}}$ the diffuse incoming irradiance, $\theta_r$ and $\varphi_r$ the reflection zenith angle and azimuth angle, $\theta_i$ and $\varphi_i$ the incident zenith angle and azimuth angle and $\mu_l$ the direction cosine of the incident solar beam on the vertical axis.

In Equation 5.1 the reflected radiance is assumed to originate from an infinitesimally small solid angle. In practice, one usually operates with instruments that have a finite field of view. For this reason it would be more appropriate to speak of the hemispherical conical reflectance factor, HCRF (Schaepman et al., 2006). Under the assumption of an isotropic bi-directional reflectance factor, however, the HCRF becomes equivalent to the HDRF. Therefore, we will refer to the HDRF in the rest of this paper.

For the experimental determination of the HDRF it is common to utilize a Spectralon panel (cf. Labsphere, New Hampshire, USA) as a proxy for a perfect Lambertian surface ($L_{\lambda, \text{Lamb}} \sim L_{\lambda, \text{Spectralon}}$). The radiance of a Lambertian surface is constant regardless of the angle from which it is viewed. Commercially available panels are never perfectly Lambertian. Departures from a perfect reference are considered by introducing a spectral correction term, $C_\lambda$, that accounts for the directional reflectance characteristics and for subunity in total reflectance (Sandmeier et al., 1998):

$$\text{HDRF}_\lambda(\theta_r, \varphi_r, \theta_i, \varphi_i) = \frac{L_\lambda(\theta_r, \varphi_r)}{\left(L_{\lambda, \text{Spectralon}}(\theta_r=0, \varphi_r=0)\right)^{C(\theta_i)}}. \quad (5.2)$$

The HDRF is a function of the wavelength and four angles (Figure 5.1). In all field experiments, HDRF datasets are measured under various solar zenith angles and azimuth angles. For easier comparability, however, the results are usually presented with respect to the solar principal plane (e.g. Steffen, 1997; Sandmeier, 2000; Painter and Dozier, 2004b). The solar principal plane is defined in such a way that the azimuth angle of the incident radiation is equal to 180° and the azimuth angle of forward scattering is equal to 0°.
Fig. 5.1: Angles and naming conventions for defining the HDRF: \( E_{\text{dir}} \) = direct incoming irradiance, \( E_{\text{diff}} \) = diffuse incoming irradiance, \( L_r \) = reflected radiance, \( \theta_i \) = solar zenith angle, \( \varphi_i \) = solar azimuth angle, \( \theta_r \) = reflection zenith angle, \( \varphi_r \) = reflection azimuth angle.

Assuming an azimuthally homogeneous surface, the number of relevant angles is reduced to three: \( \theta_i, \theta_r \) and \( \varphi_i - \varphi_r \).

The extent of the inhomogeneous reflectance of one measuring cycle can be described with the Anisotropy index, \( \text{Anix} \):

\[
\text{Anix}_\lambda = \frac{\text{max}(\text{HDRF}_\lambda)}{\text{min}(\text{HDRF}_\lambda)}.
\] (5.3)

The spectral albedo, \( \rho_\lambda \), can be calculated by integrating the HDRF\(_\lambda \) for all reflectance angles over the hemisphere and normalizing the result with \( 1/\pi \):

\[
\rho_\lambda(\theta_i, \varphi_i) = \frac{1}{\pi} \int_0^{2\pi} \int_0^{\pi/2} \text{HDRF}_\lambda(\theta_i, \varphi_i, \theta_r, \varphi_r) \times \cos(\theta_r) \sin(\theta_r) d\theta_r d\varphi_r.
\] (5.4)

A broadband albedo over a certain range of wavelength \([\lambda_1, \lambda_2]\), \( \rho_{\lambda_1}^{\lambda_2} \), can be obtained as a weighted integral of \( \rho_\lambda \):

\[
\rho_{\lambda_1}^{\lambda_2} = \frac{\int_{\lambda_1}^{\lambda_2} \rho_\lambda \ E_\lambda \downarrow d\lambda}{\int_{\lambda_1}^{\lambda_2} E_\lambda \downarrow d\lambda}.
\] (5.5)
where $E_{\lambda \downarrow}$ is the total spectral irradiance in $\lambda$, whereas the total albedo is:

$$\rho = \frac{\int_0^\infty \rho_{\lambda} E_{\lambda \downarrow} \, d\lambda}{\int_0^\infty E_{\lambda \downarrow} \, d\lambda}.$$  

(5.6)

5.3 Experiments

5.3.1 Experimental site

The experiments were carried out at the Summit Environmental Observatory, 72° 35' N, 34° 30' W, http://www.geosummit.org/ (Ohmura, 2001). At Summit, year around measurements of all the components of the energy balance are carried out since 2001 by the Institute for Atmospheric and Climate Science, Swiss Federal Institute of Technology (IACETH). These include measurements of the radiative fluxes such as longwave and shortwave incoming and reflected radiation as well as micrometeorological measurements needed to compute the turbulent exchange of sensible and latent heat. Furthermore, snow temperature is recorded at the surface and at 15 levels within the snow cover down to a depth of 15 m. Irregular snowpit work was performed in order to determine snow density and stratigraphy.

The station is situated at 3200 m a.s.l.; the atmospheric optical depth is small and the air and the snow can be assumed to be clean due to the large distance from sources of atmospheric pollutants. The site lies in the dry snow zone where no melting occurs. Nevertheless, the snow surface characteristics can vary significantly within short time periods. In the summers of 2004 and 2005, two main types of snow surfaces were observed at Summit: smooth surfaces with wind-broken small snow grains as a product of wind drift, coarse surface rime, accreted during fog events.

5.3.2 Instrumentation: IACETH Goni-Spectrometer

The utilized IACETH Goni-Spectrometer will only briefly be described. For a detailed technical description of the entire instrumentation we refer to Bourgeois et al. (2006b). Figure 5.2 shows a drawing of the goniometer with all relevant parts. In Figure 5.3 the deployment of the IACETH Goni-Spectrometer in the field is depicted. As seen in the figures, the whole assembly consists of a vertical post, a horizontal boom fixed on this post and two moving arms attached to the end of the boom. This construction allows to put the sensor at any position on the hemisphere above the target area, viewing a varying area located by the center point of the field-of-view. The 3° field-of-view foreoptic results in an elliptical footprint with a major diameter of 5.2 cm at nadir and 31 cm at 80°. The leveling of the instrumentation including the Spectralon panel is observed and adjusted before each measuring cycle. The distance between sensor and target point on the surface is set to one meter and controlled with a laser distance meter.

The IACETH Goniometer is operated with a standard field spectrometer, 'FieldSpec Pro Dual VNIR', from 'Analytical Spectral Devices', Boulder CO. The spectrometer was calibrated together with the foreoptics in 2004 before the deployment to the field. The accuracy of the instrument is better than 1%. The spectrometer covers a nominal spectrum between 350 and 1050 nm. The spectrum is measured with a 512-channel silicon
5. HDRF and Spectral Albedo of Dry Snow

The Lambertian reference used to calculate the reference reflectance was a Spectralon panel manufactured at Labsphere (information available online at http://www.labsphere.com). The factory provides spectral calibration coefficients which account for the subunity in reflectance. Additionally, we corrected the anisotropic characteristics of the Spectralon panel with the algorithm proposed by Sandmeier et al. (1998).

5.3.3 Field program

Field experiments to determine the HDRF and spectral albedo of snow were carried out in 2004 and 2005, from 27 May to 4 July and 18 June to 24 July, respectively. The time needed for a complete HDRF cycle with an angular resolution of 15° in azimuth and the viewing zenith angles of 0°, 15°, 30°, 45°, 60°, 75° and 80° is 11 minutes corresponding to a change of less than 4° in solar zenith and azimuth angle. The solar zenith angles varied between 49° and 85°. Measurements were made every one or two hours, provided that cloud coverage on the sky was less than 10%. The persistence of cloudless conditions during a measuring cycle was monitored with a second probe which allowed to take simultaneous measurements of the incoming spectral irradiance. A 2π foreoptic was mounted on the reference fiber. The reference fiber showed single broken glass threads which prevented using it for calculation purposes, but it was still useful to monitor the variation of the incoming irradiance.

The horizontal arm of the Goniometer was always left in the same azimuthal position. The actual solar azimuth was determined for each measuring cycle. Before and after each cycle a sample on the Spectralon panel was taken.
5.3. Experiments

Fig. 5.3: Setup of the IAC ETH Gonio-Spectrometer in the field in different measuring positions.

Tables 5.1 and 5.2 show all experiments of good quality, including the measuring time and the illumination geometry. In the same tables the broadband albedo, measured with two pyranometers (one upfacing, one downfacing), the integral albedo in the range 400 to 1000 nm calculated from the directional measurements, and the spectral albedos for 500, 700 and 900 nm are listed.
5. HDRF and Spectral Albedo of Dry Snow

Table 5.1: List of the experiments in 2004 with the sun geometry (θi, ϕi), the broadband albedo 280-4000 nm (ρ), the broadband albedo 400-1000 nm (ρ_{400-1000}) and the spectral albedo for 500, 700 and 900 nm (ρ_{500}, ρ_{700}, ρ_{900}).

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<th>ϕi (°)</th>
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<th>ρ_{400-1000}</th>
<th>ρ_{500}</th>
<th>ρ_{700}</th>
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</tr>
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</table>

In 2004, the observation of the snow surface characteristics was made with an aid of digital pictures and with a grid plate and a magnifying glass. In 2005, additionally, snow samples from the uppermost snow surface layer were collected. The phthalate method was applied to preserve the samples (Good, 1989). The air within the snow samples was carefully filled with dyed dimethyl phthalate and the frozen samples were transported to the Swiss Federal Institute for Snow and Avalanche Research, SLF Davos. There, the samples were shaved with a microtome and digital pictures of the cross-sections were produced. The snow grain shape, snow grain size and snow density were estimated from the binary-coded pictures.
### Table 5.2: List of the experiments in 2005 with the sun geometry ($\theta_i, \varphi_i$), the broadband albedo 280-4000 nm ($\rho$), the broadband albedo 400-1000 nm ($\rho_{1000}^{400}$) and the spectral albedo for 500, 700 and 900 nm ($\rho_{500}$, $\rho_{700}$, $\rho_{900}$).

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### 5.3.4 Data Processing, quality control and corrections

Altogether 193 complete HDRF cycles were sampled in 2004 and 2005. After a quality control, 76 cycles were retained for the analysis. The following criteria had to be fulfilled: 1) no observable changes in cloudiness and cloud distribution; 2) ideal optimization to adjust the sensitivity of the instrument’s sensor to the specific illumination and 3) no
saturation of the sensor for any wavelength range during the whole cycle.

Self shadowing of the target area by the instrument was corrected by assuming symmetry across the solar principal plane (Aoki et al., 2000), which makes it possible to replace data affected by the instrument shadow with corresponding values on the unaffected hemisphere.

5.4 Results

5.4.1 HDRF of snow

The large number of HDRF cycles, obtained under a wide range of variations in radiation and surface features, provides a good basis to investigate the HDRF in relation to solar zenith angles and snow characteristics.

Figures 5.4 and 5.5 show the HDRF datasets collected for different snow roughnesses and snow grain characteristics: a smooth surface with wind broken small snow grains (maximum diameter <0.5 mm) observed on 21 June 2005 in Figure 5.4, and a surface covered with rime, observed on 27 June 2005 in Figure 5.5.
5.4. Results

Fig. 5.4: **a)** Polar plots of the HDRF for three wavelengths (500, 700, 900nm) and two different solar zenith angles (49° and 58°), 21 June 2005. The snow surface was smooth and the snow grains small (≤0.5 mm) and broken from winddrift. The plots only extend to a view zenith of 80°. The solid contour represents HDRF=1; the dashed contours are plotted with a 0.2 interval. The number next to each plot is the calculated Anix (see Eq. 5.3). Note that the ranges of the colorbars are different for each plot. **b)** Variation of the HDRF along the solar principal plane for the same wavelengths. The HDRF in the solar principal plane goes to larger values than in the polar plots above, this because no measuring points are available there.
Fig. 5.5: a) Polar plots of the HDRF for three wavelength (500, 700, 900nm) and three different solar zenith angles (49, 58 and 85), 27 June 2005. The snow surface was covered with surface rime, the snow grains where large (>1 mm). The plots only extend to a view zenith of 80°. The solid contour represents HDRF=1; the dashed contours are plotted with a 0.2 interval (omitted for HDRF >3). The number next to each plot is the calculated Anix (see Eq. 5.3). Note that the ranges of the colorbars are different for each plot. b) Variation of the HDRF along the solar principal plane for the same wavelengths. The HDRF in the solar principal plane goes to larger values than in the polar plots above, this because no measuring points are available there. The range on the axis of the ordinate in the last plot is about 10 times as large as in the first two plots.
Persistent rime accretion resulted in cm-scale surface roughness and irregular shaped snow grains. The maximum diameter increased to \(\sim 3\) mm. The results of the experiments are displayed as polar plots, whereby the actual measuring points are marked with crosses (+). The radial distance from the center represents the view zenith angle, and the rotation about the center represents the view azimuth. The three panels on the left of both figures show the HDRF measured at a solar zenith angle of \(\theta_i=49^\circ\) for the three wavelengths 500, 700 and 900 nm. Corresponding plots for \(\theta_i=58^\circ\) are shown on the right hand side. On 27 June a measuring cycle under a solar zenith angle of 85\(^\circ\) was collected and the data are displayed in the third column of Figure 5.5. However, in this experiment the instrument was not ideally optimized in the wavelength range 550 - 600 nm and this data was not considered for further processing. Snow surface conditions at the times of the measurements shown in Figures 5.4 and 5.5 are presented in Figure 5.6. Cross sections through the uppermost 7 cm of the preserved snow samples are depicted in Figure 5.7. The pictures are made from the preserved snow samples. As can be seen in the figures the snow grains were small, regular shaped and dense packed on 21 June, while on 27 June the snow grains were larger and irregular, and the density was much lower.

Values of the HDRF for the smooth surface (Figure 5.4) are lower in the backward (<0.8) and higher (1.0-1.6) in the forward direction than for the rime covered surface (0.8-1.0 and 0.9-1.1)(Figure 5.5). As a consequence the HDRF of the smooth surface displays a higher Anix (~2.0) than the HDRF obtained for surface rime (~1.2) for solar zenith angles of 49\(^\circ\) and 58\(^\circ\). A strong forward scattering peak at the largest measured view zenith angle of 80\(^\circ\) can be seen for the smooth surface (HDRF \~1.2 at \(\theta_i=49^\circ\), HDRF \~1.7 at \(\theta_i=58^\circ\), Figure 5.4b), while the rime covered surface produces a less pronounced back scattering peak at a view zenith angle of 60\(^\circ\)-70\(^\circ\). For smaller solar zenith angles the HDRF distribution over the hemisphere is more uniform for the rime covered surface with HDRF values around 1.0.

Figures 5.4b and 5.5b illustrate the wavelength dependency of the HDRF. Generally, the values decrease with increasing wavelength. Exceptions are the forward scattering direction at large solar zenith angles where the HDRF values of longer wavelengths become larger and exceed the ones of the shorter wavelengths, and on rough surfaces where the HDRF values at large solar zenith angles are higher for all wavelengths. Datasets collected under large solar zenith angles (see Figure 5.5) reveal a very strong forward scattering with maximum HDRF values of 13 at the view zenith of 80\(^\circ\).

In all cases the HDRF distribution shows an increase in the forward direction, but depending on the surface characteristics the distribution of minima and maxima are different. Remarkable is the difference between the observations at solar zenith angles of 49\(^\circ\) and 58\(^\circ\). For the rough surface the backscattering peak moves from \(\theta_r=45^\circ\) at \(\theta_i=49^\circ\) to \(\theta_r=60^\circ\) at \(\theta_i=58^\circ\). At the reflectance zenith of 80\(^\circ\), in the forward direction, the HDRF increases \~0.5 for the smooth surface but only \~0.1 for the surface rime (see Figures 5.4b and 5.5b).
Fig. 5.6: Digital pictures of the snow surfaces on 21 June (smooth surface, roughness elements <0.5 cm) and 27 June (surface rime, roughness elements >1 cm) 2005. The white plate is the Spectralon panel used as a Lambertian reference. The reference panel has a side length of 12.7 cm and is mounted on 34 cm above the surface. Typical dimensions (not the effective grain size), of the snow grains are of max. 0.5 mm in the upper panel and of max. 3 mm in the lower panel (see text).

In Figure 5.8 the Anix for all experiments is shown as a function of the solar zenith angle. Despite the large variations at individual solar zenith angles, which can be explained with the varying surface characteristics, it remains approximately constant (∼1.5) up to $\theta_i \sim 65^\circ$, increasing with the solar zenith angle above this threshold. At $\theta_i \sim 75^\circ$ the Anix reaches a value of ∼3.0.
Fig. 5.7: The figure shows the binary representation of photographs from snow samples collected on 21 June 2005 and 27 June 2005. The samples were preserved with the phthahlate method (Good, 1989). The snow grains are black colored and the air in between is white. Snow grains in the top 1 cm of the profile on the left are homogeneous with a max. diameter of 0.5 mm, whereas snow grains in the uppermost 2 cm on the profile on the right are flat and irregular with a max. diameter of 3 mm. The roughness of the surface on July 21 was estimated at ∼1 cm.

Fig. 5.8: Median, minimum and maximum value of the Anix for 500, 700 and 900 nm as a function of the solar zenith angle. The number of samples is ∼4 for θ_i < 60°, but only of one or two for θ_i > 60°.
5.4.2 Reflectance in the solar principal plane

In Figure 5.9 the mean HDRF values along the solar principal plane are compared for various solar zenith angle ranges.

Fig. 5.9: Median, minimum and maximum values of the HDRF in solar principal plane for four ranges of solar zenith angles. The wavelength is 500 nm in the upper four panels and 900 nm in the lower four panels.
The upper four panels represent the wavelength of 500 nm and the lower four panels the wavelength of 900 nm. The medians with the minimum and maximum values of all data from 2005 passing the quality control are plotted. The experiments have been grouped in four classes according to their illumination geometry: a) \( 48^\circ < \theta_i < 50^\circ \), b) \( 50^\circ < \theta_i \leq 55^\circ \), c) \( 55^\circ < \theta_i < 60^\circ \), d) \( 60^\circ < \theta_i \leq 70^\circ \). Experiments with larger solar zenith angles are rare due to experimental difficulties: the radiation intensities become very weak and the signal to noise ratio decreases.

The graphs of the solar principal plane clearly demonstrate that at all times the forward scattering is stronger than the backward scattering. A large difference can be seen for the data collected in the evening hours where the forward scattering becomes very pronounced and the function over the hemisphere becomes bowl shaped with a minimum between -30° and 30° for the view zenith angle.

### 5.4.3 Spectral snow albedo

Figure 5.10 shows two spectral albedo datasets measured on the same day but one at a solar zenith angle of 49° and the other at 58°. The surface was covered with rime resulting in a cm-scale surface roughness (see Figure 5.6b). The spectral albedo was calculated according to Equation 5.4. Thereby the measured HDRF was first interpolated on a 2° spherical grid using a Delaunay interpolation (Renka, 1984). The albedo shows values higher than 0.9 for wavelengths smaller than 700 nm and decreases with increasing wavelength, with two local maxima at 840 and 940 nm and two local minima at 800 and 910 nm. The spectral albedo increases with increasing solar zenith angle.
5.4.4 Broadband albedo

In Figure 5.11a the albedo computed according to Equation 5.6 for the wavelength range 400 to 1000 nm is compared to the broadband albedo for the range 280 to 4000 nm measured with two back-to-back pyranometers. The albedo data are plotted as a function of the local solar time. The dashed line represents a fit through the goniometer data and the solid line represents the mean albedo of the clear sky days measured operationally with the pyranometers.

Values of the broadband albedo in the range [400, 1000] nm are systematically higher than those of the broadband albedo [280, 4000] nm. Furthermore the derived broadband
albedo from directional measurements exhibits a stronger diurnal cycle than the broadband albedo [280, 4000] nm. The graphs are almost symmetrical about solar noon. The weak asymmetry can be explained either with a not perfectly horizontal surface (Ohmura, 1980; Grenfell et al., 1994) or snow metamorphosis in the course of the day.

Figure 5.11b shows the values of a modeled albedo as a function of the integration bandwidth. A standard irradiance curve for a subarctic summer atmosphere (altitude=3200 m, T_{surface}=287\degree) has been calculated with the SBDART model (Ricchiazzi et al., 1998). The broadband albedo was then calculated by integrating the built-in spectral snow albedo model. The lower limit of the integral was kept at 280 nm and the upper limit was progressively increased to 4 \( \mu \)m. The figure shows that with increasing bandwidth the value of the albedo first increases with a maximum at 550 nm and then decreases slowly reaching 0.7 at 4 \( \mu \)m. Overall, these calculation suggest that a difference of 0.1 between the broadband albedo [280, 4000] nm and the value for the range [400, 1000] nm can be explained by the spectral distribution of the snow albedo and the solar irradiance.
5.5 Discussion

The dominating factor influencing the HDRF is the solar zenith angle (see Figure 5.5). This fact supports the approach of Warren et al. (1998) who proposes a parametrization for the HDRF using the solar zenith angle and the view zenith angle as variables. Nevertheless, the experimental results show that the physical characteristics of the snow surface have to be considered for an accurate HDRF parametrization for zenith angles < 65°.

The examples presented in this study (see Figures 5.4 and 5.5) illustrate that the diurnal variation of the snow HDRF observed during the measuring campaigns follows a general pattern. At solar noon the HDRF pattern shows a flat distribution with a peak in forward scattering. A strong increase of the forward scattering with increasing solar zenith angle is observed in most measurements. An explanation for the increasing forward scattering with increasing solar zenith is that the direct beam impinging with a small angle on the surface penetrates less deep into the snow and undergoes less scattering than a vertical beam. In the latter case many scattering events are required for the beam to be redirected upward and eventually escape the snowpack. The photons are then distributed more uniformly back into the upward hemisphere.

Larger wavelengths generally show smaller HDRF values because their absorption in the snow is stronger (Wiscombe and Warren, 1980; Bourgeois, 2002). However, for large solar zenith angles and surface rime the reverse was observed (see Figure 5.5b).

Our measurements revealed that cm-scale snow surface characteristics have a substantial influence on the reflectance characteristics (see Figures 5.4 and 5.5). On a macroscopic scale (diameter >10 cm) both surfaces look homogeneous, however, the snow surface with rime accretion exhibits a lower density within the uppermost 2 cm and shows significantly different snow grain shapes. On the smooth surface the direct beam undergoes only few scattering events within the uppermost layer and the photons are predominantly scattered in the forward direction. On the rime covered surface, for θi < 70°, the reflectance is rather uniformly due to more scattering events within the first two cm of the snow cover.

In this study the spectral albedos are derived from the HDRF measurements according to Equation 5.4. Considering the results of earlier studies (Choudhury and Chang, 1979; Wiscombe and Warren, 1980) and the variability of the HDRF shown above, it can be expected that the spectral albedo varies in relation with the sun zenith angle, the snow surface characteristics and the impurity content. The observed spectral albedos showed remarkably low values in the visible range (see Tables 5.1 and 5.2). This suggests the presence of absorbing impurities in the snow (Warren and Wiscombe, 1980). However, in this study we did not measure the amount of impurities.
5.6 Conclusions

A comprehensive dataset of experimental HDRF measurements is presented. Field work has been carried out on the ice sheet of Greenland in the dry snow zone where no melting occurs. The dataset represents summer conditions only. For a climatology we would need additional data for spring and autumn (different solar illumination and snow surface characteristics). The HDRF was measured with a specially developed instrumentation, the IACETH Gonio-Spectrometer. The collected datasets show a strong variation of HDRF with the solar zenith angle, increasing at higher wavelength. The study further reveals that the cm-scale snow surface roughness of the uppermost snow layer has a strong influence on the HDRF. Rime accretion on the surface produces extended irregular snow crystals and a weak backscattering peak was observed for $\theta_i < 70^\circ$. However for large solar zenith angles specular reflectance may occur and lead to a very large forward scattering. Smooth surfaces with wind broken small snow grains always revealed strong forward scattering peaks.

The results of this study show that the knowledge of the reflectance function over the hemisphere and its dependence on the solar geometry is crucial for an accurate albedo retrieval from remote sensing platforms. Calculating a broadband albedo from only one measurement and assuming an isotropic reflectance over the hemisphere can lead to major errors, depending on the measuring angle and the illumination angle.

The spectral albedo derived from HDRF datasets shows some inconsistencies with previous measurements and models. There are two possible explanations: numerical issues of the applied interpolation algorithm and strongly absorbing impurities in the snow.

In this work we assumed a perfectly even and horizontal surface. Earlier studies showed that neglecting the tilt of the surface can lead to significant errors. Therefore in the future it will be desirable to find ways for actually measuring the tilt and to assess how the error increases with tilt. This could however be a major challenge. Furthermore, future studies should stress the investigation of the snow stratigraphy to better explain the observed reflectance features. Measurements of HDRFs on slightly tilted and on not ideally flat surfaces would help to finally incorporate HDRF models into retrieval algorithms for rough terrain.

As already mentioned in Section 5.2, given the $3^\circ$ field-of-view foreoptic, the quantity measured in this study was the HCRF rather than the HDRF. Still, we referred to the HDRF under the implicit assumption that the bi-directional reflectance factor was isotropic within the solid angle sampled by the instrument. It was shown in Bourgeois et al. (2006b) that the corresponding footprint area varied in maximum diameter from 5 to 31 cm, depending on the viewing position. In most experiments carried out for the present study, the surface properties of the snow could be considered as uniform within this footprint area. This qualitatively justifies the assumption of an isotropic bi-directional reflection factor. Nevertheless, an accurate verification of this assumption should be included in future work.
Acknowledgment

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Chapter 6

General Conclusions and Outlook

6.1 Conclusion

In this thesis experimental studies measuring the radiative properties of snow are presented. This chapter contains a brief summary of the main subjects and possible directions for future research are discussed.

The significance of snow radiative and in particular reflectance properties have given rise to many theoretical and experimental studies. A short historical overview on these studies is given in Chapter 1. The experimental site of this study, the Summit Environmental Observatory on the ice sheet of Greenland and the research programme of the IAC at Summit are presented in Chapter 2. Theoretical aspects of snow radiative processes and in particular reflectance properties have been studied extensively during the last three decades.

Chapter 3 presents a detailed literature review on experimental and theoretical works investigating on the optical snow properties, the broadband reflectance of snow, the spectral reflectance of snow and the directional reflectance of snow. The chapter is complemented with the results of my own experiments carried out at Summit in 2001, measuring the penetration of solar radiation into the snow cover. Results of the 2001 study lead the author to the main objectives of the present thesis. The largest part of the solar incoming radiation is directly reflected back to the atmosphere. But the magnitude of the reflection is strongly dependent on the snow surface properties. The highest portion of the absorbed radiation is absorbed at the atmosphere-snow interface and only a very small portion penetrates deeper into the snow cover. The penetrating process is highly dependent on the wavelength and the snow properties of the uppermost 10 cm of the snow cover.

To explore the directional and spectral reflectance of the snow the IACETH Goni-Spectrometer has been developed (Chapter 4). The instrument fulfills the requirements of measuring the reflected radiation over the hemisphere with high angular accuracy within a short acquisition time: 11 minutes for covering the entire hemisphere at 15° grid-spacing. Investigations showed a pointing accuracy of the sensor of ±1 cm. Depending on the view zenith the footprint varied from a circle with a diameter of 5.2 cm at nadir to an ellipse with a major diameter of 31 cm at 80° view zenith. Shadowing of the target by the instrument when it aligns with the sun was accounted by replacing the affected measurements with the corresponding values from the other side of the solar principal plane. First results from the 2004 measuring campaign are presented in Chapter 4. The results show the wavelength dependency, the solar zenith angle dependency but also the
influence of the snow surface properties.

A complete overview on the collected HDRF datasets from the two measuring campaigns in 2004 and 2005 at Summit is presented in Chapter 5. In order to get a better picture of the snow properties, snow samples from the uppermost 7 cm were collected in 2005 and analyzed with respect to the snow grain size and snow grain shape. Measurements were carried out for solar zenith angles in the range of 49° to 85°. The strong forward scattering of snow surfaces increases by one order of magnitude at large solar zenith angles. At moderate solar zenith angles the forward scattering was more pronounced on surfaces with a small surface roughness and small snow grains while cm-scale surface roughness due to rime accretion created small back scattering peaks in the solar principal plane. With an interpolation and integration algorithm the spectral reflectance was calculated from the directional measurements. The spectral signatures varied with solar zenith angle and snow properties, however, the variation of the spectral signature could not be explained satisfyingly. A comparison of the broadband albedo from directional measurements to broadband albedo measured with pyranometers revealed the strong dependence of albedo on the integrated wavelength range. A smaller wavelength range produces a stronger daily cycle with higher albedos in the morning and evening hours than full range measurements.

6.2 Outlook

Based on the foregoing conclusions, a series of new questions emerge. Future research could address the following aspects:

- **HDRF measurements on different snow types and melting snow**: The snow at Summit showed some variability. However, under different climatological conditions the snow surfaces exhibit different characteristics. For example snow in the ablation zone undergoes melting processes. The melting and refreezing processes alter a surface and the physical properties of a snow surface drastically.

- **HDRF measurements on tilted surfaces**: The homogenous conditions at Summit with its flat and horizontal surface is not representative for a large part of the Earth’s snow covered surface. The applicability of directional reflectances corrected for sloping surfaces has to be investigated.

- **Influence of clouds**: In this study only measurements under clear skies were performed. This is representative for only a small number of real conditions. To investigate the influence of clouds on the directional reflectance will be a challenge since the radiative properties of clouds already provide a complex and highly varying boundary condition for incoming radiation.

- **Calculating the HDRF with the measured incoming irradiance as reference**: The directional reflectance was calculated with the aid of a Lambertian Spectralon panel. However, the panel is known not to be a perfect Lambertian reflector. Calculating the reflectance with the help of measured irradiance would allow to minimize this effect.
6.2. Outlook

- **HDRF parameterizations:** A large amount of data has been collected for this study. But a parametrization of the HDRF was not performed. The results of the measurements showed such a high variability which could not be explained by the simultaneously collected snow metadata. The physical snow parameters (surface roughness, grains size, grain shape, impurities, stratigraphy...) need to be measured in more detail. Not only a precise knowledge of the snow grain sizes and shapes would be helpful but also the surface roughness and the snow stratigraphy of the uppermost 20 cm should be investigated. It further can be assumed that measuring a broader wavelength range would show more variations in the spectral signatures for various snow surfaces. A simultaneous measurement of all these properties for the major snow types would provide a solid foundation for future development of snow radiative transfer models.

- **Investigations on the spectral albedo derived from directional measurements:** The hemispherically applied interpolation algorithm is assumed to produce artifacts and therefore the spectral signature derived from the directional measurements are difficult to interpret. Research on the interpolation has to be included in future studies.

- **Comparison of atmospherically corrected directional surface-based measurement with satellite data:** Presently the available remotely sensed data product are often composites of several days. They are not suited for validation purposes since snow surfaces vary strongly with daytime. Eventually satellite data will become available with high spatial and temporal resolution and retrieving algorithms of remotely sensed data can be validated with surface-based data.

This work also points to further issues and questions of more general scope:

- **General applicable HDRF parametrization:** The final goal would be to model the HDRF from a limited number of measurements collected with satellites for known illumination geometries. Together with elevation models, HDRF values then have to be corrected to the appropriate functions. And finally broadband albedos could be retrieved and used for climate modeling.

In conclusion, this study has shed more light on the mysteries of the radiative and reflectance properties of snow. This has become possible with the development of new instruments like spectrometers and goniometers to carry out field measurements. Nevertheless, the study has also shown that there are still many open questions. More research effort should be invested towards understanding the reflectance properties of various snow surfaces, and towards developing model formulations for the snow directional reflectance. This is especially important as remote sensing datasets will improve in quality and become more readily available. It is hoped that the present thesis will contribute towards comprehending an immensely complex reality.
Appendix A

Spectral Albedo of Various Snow Types

A.1 Introduction

This section presents an experimental study with the goal to get a better understanding of the effect of a snow type on the spectral reflectance. It is a cooperative work of the Swiss Federal Institute for Snow and Avalanche Research, SLF Davos, the Institute of Environmental Geosciences, University of Basel and the Institute for Atmospheric and Climate Science, ETH, Zürich. The spectral reflectance of short wave radiation of five snow samples was measured and then modeled with a beam tracing model (BTM). The three-dimensional structure of the snow samples was analyzed with a number of methods to derive parameters for the BTM model and to interpret the measured reflectances. The driver for the study was the empirical correlation found by Matzl and Schneebeli (2006) between the specific surface area (SSA) of snow samples and the reflectance in the near infrared. The result of the mentioned study shows increasing reflectance with an exponentially increasing SSA. The measurements did not provide information whether values deviating from the fit are only due to measurement imprecision or also conditioned by other structural parameters. To address these questions we measured the spectral reflectance of 5 different snow samples in the laboratory under controlled temperature and illumination conditions. With an X-ray micro-tomograph the snow structures were recorded as digital images of the cross sections of the natural snow samples. From these cross sections the SSA was estimated stereologically and the reflectance was simulated with the BTM. Here we only show the results of the reflectance measurements and the correlation with the SSA. The BTM, originally developed to simulate reflectances of soils Bänninger et al. (2006) did not perform in a satisfying way and needs more work.

A.2 Material and methods

The snow samples for this study were collected in the field in the vicinity of the SLF at Davos and stored at different temperatures in Styrofoam boxes with dimensions of 30 x 30 x 30 cm. For the reflectance measurements, the surface of the snow samples was prepared by cutting a thin layer from the block with a sharp metal plate. The reflectance was then recorded with a standard field spectrometer 'FieldSpec Pro Dual VNIR', from 'Analytical Spectral Devices', Boulder CO. The spectrometer covers a nominal spectrum between 350 and 1050 nm. The spectrum is measured with a 512-channel silicon photo-diode array. For the measurements we attached a 3° field-of-view foreoptic to the fibre measuring the
radiance. Each sample was scanned 15 times to minimize measuring errors. The light source was a quartz lamp mounted to an Ulbricht sphere. Light leaving the circular opening of the Ulbricht sphere is nearly 100% diffuse. The samples were illuminated with the entire wavelength spectrum of the light source. Before the reflectance of each sample was measured, the radiance of a white reference, a Spectralon panel (manufactured at Labsphere, North Sutton NH, USA) was measured. The reflectance was then calculated according to the following equation:

\[ R_\lambda = \frac{L_{\lambda,\text{snow}}}{L_{\lambda,\text{Spectralon}}} \quad [-] \]  

where \( R_\lambda \) denotes the spectral reflectance and \( L_\lambda \) the spectral radiance from the snow and the Spectralon, respectively. The complete experimental setup for the reflectance measurements is depicted in Figure A.1.

![Experimental setup for the spectral reflectance measurements of snow in the laboratory.](image)

We tested the homogeneity of the collected snow with the snow micropen Pielmeier (2003), both horizontally and vertically. Then we extracted samples by cutting out cylinders horizontally. All samples were measured in a X-ray micro-tomograph (Scanco micro-CT 80). We used the SSA (specific surface area) to describe the snow structure. The SSA was estimated for vertical surface sections using model-based stereology (Baddeley and Jensen, 2005). In the following the SSA is defined as surface-to-volume ratio [mm\(^{-1}\)]. The SSA has been used for several years as an optical equivalent sphere to describe optical properties of snow (Grenfell and Warren, 1999). This property has been used to parameterize structures for radiative transfer modeling of snow (Dozier et al., 1988). Alongside its use for the optical description of snow, the SSA is an important parameter for describing the structural size and the geometry of sintered media.
A.3 Results

Table A.1 characterizes the five snow samples (short name, storage temperature, density, SSA, reflectance at 500 nm and reflectance at 900 nm).

<table>
<thead>
<tr>
<th>short name</th>
<th>storage temp.[°C]</th>
<th>density [kg/m³]</th>
<th>SSA[mm⁻¹]</th>
<th>refl. at 500</th>
<th>refl. at 900</th>
</tr>
</thead>
<tbody>
<tr>
<td>fs</td>
<td>-50 for three days</td>
<td>110</td>
<td>56.22</td>
<td>0.954</td>
<td>0.724</td>
</tr>
<tr>
<td>m1</td>
<td>-17 for three days</td>
<td>150</td>
<td>36.13</td>
<td>0.931</td>
<td>0.708</td>
</tr>
<tr>
<td>m2</td>
<td>-3 for three days</td>
<td>194</td>
<td>26.29</td>
<td>0.875</td>
<td>0.664</td>
</tr>
<tr>
<td>dh</td>
<td>-50</td>
<td>305</td>
<td>8.38</td>
<td>0.849</td>
<td>0.524</td>
</tr>
<tr>
<td>ws</td>
<td>-50</td>
<td>535</td>
<td>5.11</td>
<td>0.731</td>
<td>0.393</td>
</tr>
</tbody>
</table>

Table A.1: Overview on the studied snow samples collected in Davos.

-fs: fresh snow, collected on 20 February 2006 the morning after precipitation and stored at -50°C
-m1: equitemperature metamorphed snow, collected on 20 February 2006 and stored at -17°C
-m2: equitemperature metamorphed snow, collected on 20 February 2006 and stored at -3°C
-ws: wet snow, collected just before the laboratory reflectance measurements on 28 February 2006
-dh: depth hoar, collected just before the laboratory reflectance measurements on 24 February 2006

Figure A.2 shows for each snow sample a photograph of a typical snow grain and a cross section recorded with the tomograph. In Figure A.3 the measured spectra of the reflectances are plotted.

A.4 Discussion

Figure A.4 shows that the reflectance decreases as the snow density increases. The values remain above 0.83 in the visible range for the samples fs, m1, m2, and dh. The ws, however exhibits in the visible range values of ~0.7. In the near infrared the reflectances decrease more or less equally for all samples. The characteristic local minima for snow reflectance at the wavelengths around 800, 900 and 1030 nm are clearly visible. In Figure A.4 the evaluated SSA of the five snow samples are plotted against the measured reflectances for three wavelengths. The SSA varies in a wide range, with values between 56.22 mm⁻¹ for the fresh snow and 5.11 mm⁻¹ for the wet snow. The reflectances decrease together with the decreasing SSAs and this for all snow samples and wavelengths with about the same strength.

A.5 Conclusion

This laboratory study shows and quantifies the growth of snow grains and the increase of the density as snow ages (see Figures A.2 and A.4). Furthermore, the measurements reveal a decrease of the reflectance with increasing SSA. However, the theory (for example
Fig. A.2: On the left panel photos of typical snow grains are depicted for the five snow samples. The right panel shows tomographed cross sections of the snow samples where white represents the air phase and black the snow particles. fs stands for fresh snow, m1 for the same snow type but stored at -17°C, m2 for the same snow type again but stored at -3°C, dh for depth hoar, ws for wet snow.
Wiscombe and Warren (1980) states that the reflectance in the visible is not dependent on the grains size but strongly depends on the impurities (Warren and Wiscombe (1980)). We assume therefore that the wet snow was contaminated. This seems plausible since the village of Davos is a source of pollution and the wet snow was exposed in the open space for several days while the other samples were collected just after a snow fall or only few days later. For a realistic interpretation of the measured reflectances and also for
the simulation with the beam tracing model it is mandatory to measure the amount of impurities in the snow.
Appendix B

Pictures of the Used Instrumentation

Fig. B.1: Spectralon panel from Labsphere, North Sutton NH, USA.

Fig. B.2: 'FieldSpec Pro Dual VNIR', from 'Analytical Spectral Devices', Boulder CO, USA.
Fig. B.3: Suitcases of the Goni-Spectrometer: the smaller one in the upper picture contains the delicate parts and the other the large parts.
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List of Acronyms and Abbreviations

α  total reflectivity of the surface (=albedo) -
θ  zenith angle °
φ  azimuth angle °
κ  extinction coefficient cm⁻¹
λ  wavelength nm
ρ  snow density g/cm³
ρₐ  spectral albedo -
ω  solid angle sr

a₀  surface absorbtivity of the snow cover -
Anix  anisotropy index -
BRDF  Bidirectional Reflectance Factor -
BTM  beam tracing model -
d  snow grain size mm
E  radiative Flux W/m²
GL  global radiation W/m²
HCRF  Hemispherical Conical Reflectance Factor -
HDRF  Hemispherical Directional Reflectance Factor -
k  absorption coefficient cm⁻¹
L  radiance W/m²/sr
LT  local time
PAR  photosynthetically active radiation
Q  direct solar radiation W/m²
q  diffuse sky radiation W/m²
r₀  surface reflectivity of the snow cover -
r  scattering coefficient cm⁻¹
rₛ  reflectivity in the snow cover -
rₛ₀  reflectivity at the upper boundary of the inner side of the snow cover -
RF  radiative flux W/m²
S↓  downward radiative flux in the snow cover W/m²
S↑  upward radiative flux in the snow cover W/m²
S₀↓  downward radiative flux at the upper boundary of the inner side of the snow cover W/m²
S₀↑  upward radiative flux at the upper boundary of the inner side of the snow cover W/m²
SSA  specific surface area mm⁻¹
UTC  united time coordinated
"Aput": A Tribute to the Inuit Language

agiuppinniq  Schneebank
anamana     Zwischenraum zwischen 2 Schneewehen
aniu        Schnee zum Wasser machen
aniuvak     Schnee an der Seite eines Abhangs
anniu       fallender Schnee
api         Schneegrund
apijuq      mit Schnee bedeckt
apusiniq    Schneewewe
aput        Schnee
aputi       Schnee auf dem Boden
aputitaq    Schneefleck an einem Berg
aqilluqaq   weicher Neuschnee
aquillqaaq  wehender, weicher Schnee
aumannaq    Schnee, der auf dem Boden schmilzt
aumannaqtuq sehr weicher Schnee
auviq       Schneeblock
iglu        Schneehütte
igluvigaq   benutzter Iglu
illusaq     Schnee, der benutzt werden kann, um einen Iglu zu bauen
imalik      Schneeeregen
isiriartaq  gelblicher oder rötlicher fallender Schnee
kaioglaq    vom Wind scharf geschliffene Schneeeoberfläche
kalutoganiq pfeilförmige Schneewewe
kanevvluk   feine Schnee-/Regen-Partikel
katakartanaq harte Kruste auf dem Schnee, die eine Fußspur hinterläßt
kavirisirlaq Schnee, der durch Regen und Frost eine rauhe Oberfläche bekommen hat
kimoaqtruk  Schneewewe
kinirtaq    feuchter, kompakter Schnee
mangiggal   harter Schnee
mangokpok   wässriger Schnee
mannguq     schmelzender Schnee
mapsuk      überhängende Schneewewe
masak       nasser, fallender Schnee
matsaaq   halb geschmolzener Schnee auf dem Boden
maujaq   weicher Schnee
minguliq   feine Schicht aus Pulverschnee
muruaneq   weicher, tiefer Schnee
natatgonaq   rauhe Oberfläche aus großen Schneepartikeln
natiruvaaq   feiner Schnee, der vom Wind getragen wird
natiruvingniq   wehender Schnee
natquik   wehende Schneepartikel
navcaq   zusammenbrechende Schneeformation
nevluq   klebrige Schneeklumpen
nittaalaqqat    harte Schneekörner
nittaalaq    mit Schnee angefüllte Luft
nutaryuk   frischer Schnee
pilrturiqiq    dünneste Schneedecke auf einem Gegenstand
piqtuq   ”geblasener” Schnee
pirsuq   Schneesturm
pirta   Blizzard
pirtpag   aufziehender Blizzard
pukak   erste Schneeschicht
putsinniq   nasser Schnee auf der Eisoberfläche
qali   Schnee in den Ästen der Bäume
qamaniq   schüsselförmige Vertiefung unter einem Baum
qanik   (leise) fallender Schnee
qanik   Schneeflocke
qanir    schneien
qanisqineq    Schneewirbel auf dem Wasser
qanittak   kürzlich gefallener Schnee
qanniatalaaq   langsam fallender Schnee
qanniapaluq   sehr heller fallender Schnee
qannirquq   Schneewetter
qannizniq   Schneefall
qengaruk   Schneewall
qetrar   Kruste auf gefallenem Schnee
quinu   dicker matschiger Schnee
quinuq   verrotteter Schnee / Schneematsch auf dem Meer
qiqumaaq   Schnee mit gefrorener Oberfläche
qirsuqaaq   erneut gefrorener Schnee
quinzhee   Schneeschutz
quumugjuit   Schneewehe
siqoq   wehender Schnee
siqoqtoaq   Kruste auf Schnee, der in der Sonne geschmolzen ist
sirmiq   geschmolzener Schnee, der für einen Iglu benutzt wurde
sitilluqaaq   Anhäufung von hartem Schnee
sullarniq   Schnee, der in einen Durchgang geweht wurde
tumarinyiq   gekräuselte Schneewehe
upsik   vom Wind ”geschlagener” Schnee
utvak   Schneeblock
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